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LIMESTONE WALLS OF OKINAWA

By D. E. Flint, G. Corwin, M. G. Dings, W. P. Fuller, F. S. MacNeil, and R. A. Saplis

Abstract

Wall-like ridges of limestone that stand well above the surrounding terrain are an interesting phenomenon on the island of Okinawa. These ridges rim a variety of topographic features, but all are believed to represent the same formative processes. Rimming ridges or walls occur along the banks of streams crossing areas of limestone, along the upthrown sides of faults, around a sink hole, and along terrace edges.

The limestone walls of Okinawa are believed to have formed by the cementation, or casehardening, of steep exposures of poorly consolidated limestone, followed by differential erosion which brings the cemented zones into relief. Cementation occurs wherever poorly consolidated limestone is exposed to alternate wetting and drying. It is best developed, however, on steeply sloping exposures where no soil accumulates to hinder drying. Differential erosion results in part from slower solution of cemented limestone, due to its greater density, and in part from accelerated solution in the surrounding soil-covered areas.

Rampart walls along the seaward edges of terraces have been described from many Pacific islands and are present on many others. They are believed to be comparable to the walls of Okinawa and all are believed to have been formed under similar conditions.

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Introduction

Acknowledgments

The island of Okinawa was mapped geologically by D. E. Flint, G. Corwin, M. C. Dings, F. S. MacNeil, and R. A. Saplis, of the U. S. Geological Survey, aided by W. P. Fuller, a member of the Natural Resources Section of General Headquarters, Supreme Commander for the Allied Powers. The work was done for the Corps of Engineers, U. S. Army, as a part of the Post Hostilities Island Mapping Program,

administered by the Office of the Engineer, General Headquarters, Far East Command.

The members of the party mapped individual areas, and the problems that arose were discussed by all. Thus, as the geologic mapping progressed, the various occurrences of limestone walls were found and discussed, and a theory was evolved to explain their origin. Therefore, the observations and ideas in this paper represent the contributions of all the authors; the writing is by D. E. Flint.

Many of the examples of limestone walls, and a statement of the theory of their origin were included in an unpublished thesis submitted to the University of Texas (1948) by W. P. Fuller. The topographic base maps were generalized from topographic sheets prepared by the 64th Engineer Base Topographic Battalion for the Engineer, General Headquarters, Far East Command.

Geology of Okinawa

General description.—Okinawa is 67 miles long and 2 to 15 miles wide. It may be divided into two provinces on the basis of physiography and geology. The northern province, comprising about three quarters of the island, has a basement of Paleozoic sedimentary rocks that have been deformed, intruded, and slightly metamorphosed. The Paleozoic basement stands as an elevated mass that has been carved into a high backbone ridge flanked by wide erosional marine terraces that are now deeply dissected. Many of the terraces may once have been covered by Pliocene and Pleistocene limestone, but now such limestones occur only locally. The southern province, comprising about one quarter of the island, has a basement of compact, gray, silty clay interbedded with silty sands, all of Miocene age. The entire southern province was covered with a blanket of Pliocene and Pleistocene limestones. Marine erosion has terraced the island and together with subaerial erosion has removed much of the limestone cover, exposing clay and sand over much of the area.

The rimming walls of limestone occur only in areas of Pliocene and Pleistocene limestone. The nature of the underlying material is unimportant to this study, and therefore only the lithology and erosional characteristics of the limestones will be considered in detail.

Lithology of limestone.—The Pliocene limestone is thought to have been deposited on an insular shelf as either a bank or a lagoonal deposit. There are no evidences of barrier reefs in the limestone, so if it is a lagoonal deposit the encircling reefs have been removed by erosion. It consists of whole and fragmental shells and tests of lime-secreting organisms mixed with much finely divided lime that may be a chemical precipitate. Fossils of mollusks, Foraminifera, and algae are plentiful, but coral fragments are rare. Near outcrops of Paleozoic rocks the limy material is mixed with much noncalcareous detritus derived from the land. As a result, all gradations are found between gravel and sand, sandy and gravelly limestone, and almost pure limestone. The uppermost zone of this sequence is a thick layer rich in spherical to ellipsoidal algal nodules.

The Pleistocene limestone is typically a reef deposit. The bulk of the unit is rubbly, detrital, rather pure limestone, rich in whole and fragmental fossils of mollusks, corals, algae, and Foraminifera. Locally corals and algae in position of growth indicate the location of the reef which was the source of the limy fragments. Only a minor amount of noncalcareous detritus is found in the formation, and this is in the immediate vicnity of outcrops of Paleozoic rocks.

Secondary cementation.—All these limestones were deposited as poorly cemented, porous, permeable aggregates of discrete particles of lime. All the rock naturally exposed, however, is hard and well cemented; only fresh material found in deep cuts or other excavations remains poorly lithified. Upon exposure to the weather the poorly lithified material soon becomes cemented as is shown in old quarry faces and road cuts. Cementation is not limited to exposures on the ground surface, because fissures, caves, joints, and fault planes also are bordered by hardened zones. The cementation seems to occur by recrystallization and by the deposition of secondary calcite between the original particles. The secondary calcite binds the particles and fills in some of the interstices, producing a hard, but still porous limestone. Certain types of limestone seem to become

FIGURE 1.—

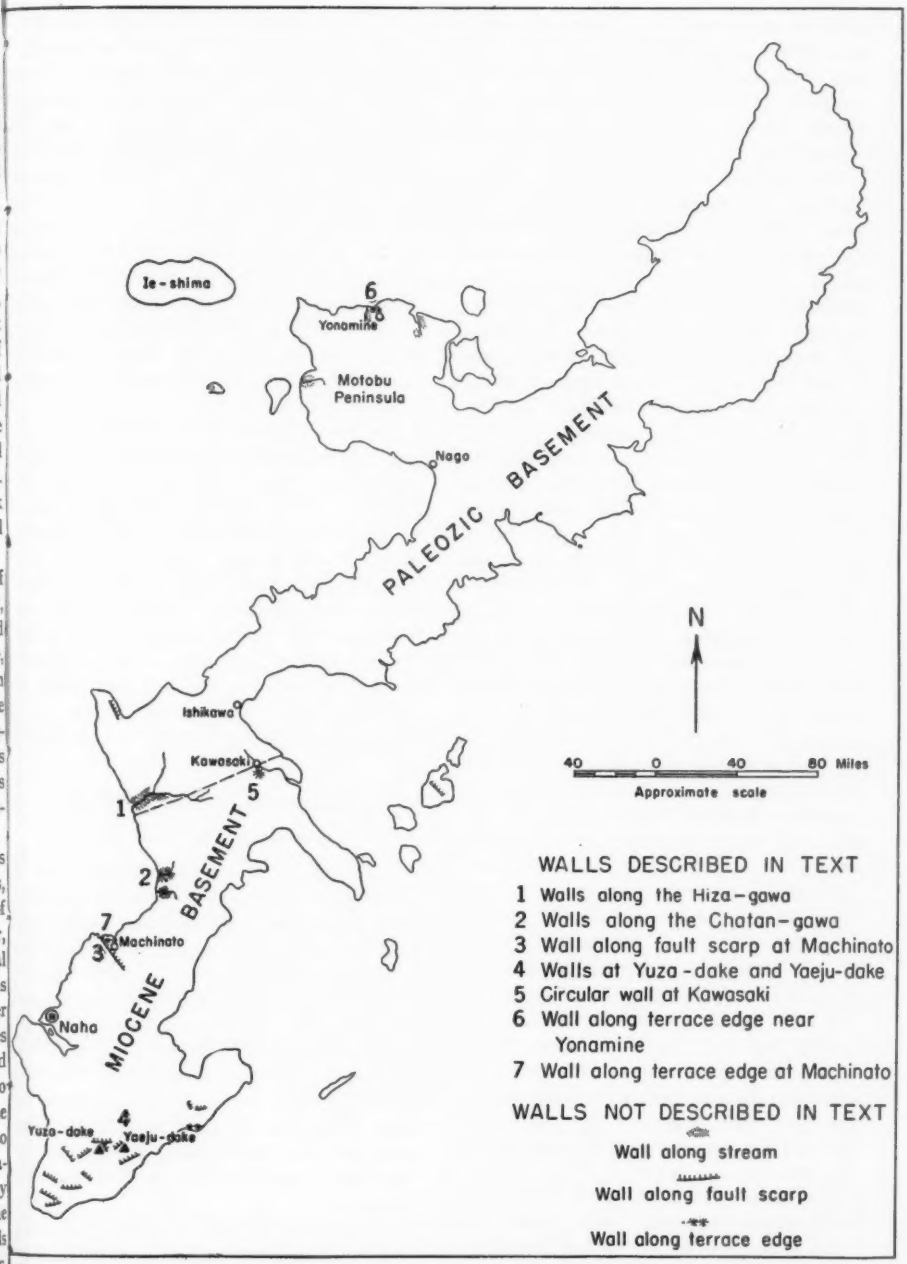

FIGURE 1.—MAP OF OKINAWA SHOWING THE LOCATION OF THE MORE PROMINENT LIMESTONE WALLS

cemented more rapidly and more thoroughly than others. Generally, the purer limestones, those with many calcareous fines, and those originally rich in aragonite are cemented more rapidly. Alternate wetting and drying aids the cementation.

This property of self-cementation or case-hardening is common in detrital organic limestones and has been utilized widely. Inhabitants of regions where such limestones occur make use of the property in quarrying. Blocks of rock are cut and shaped while the material is soft and easily worked. The rock then hardens and continues to harden after it is emplaced in wall or chimney. Military engineers utilized the self-cementing properties of limestone in the construction of airfields and roads throughout the western Pacific. Limy aggregate obtained from quarries or beaches was laid down as a surfacing material, and rolled. A hard surface developed very rapidly on the compacted material by natural cementation. The engineers found that the cementation was most rapid when the airfield or road was alternately sprinkled and allowed to dry.

Erosion of limestone.—Erosion of the limestone areas is largely by solution; mechanical erosion is almost negligible. The rock is so porous and permeable that there is little or no surface runoff. Rain falling on areas of limestone sinks into the rocks, moving through the countless interstices downward to the water table, where it begins a lateral movement down the hydraulic gradient. If the limestone is underlain by basement rock above the water table, as is generally true on Okinawa, the subsurface water runs off down the surface of the relatively impermeable basement. Streams develop along that surface in much the same manner as surface streams. These underground streams form large caverns floored in basement rock and roofed with limestone. Surface streams that enter areas of limestone either disappear into underground channels or cross the area in open canyons. The canyons are floored by the underlying basement rock and have steep to vertical sides of hardened limestone. Such canyons are believed to have formed by the unroofing of underground channels.

As meteoric water moves downward through limestone it dissolves and carries away calcium carbonate, leaving behind any insoluble materials which accumulate as residual deposits. Because the limestones differ greatly in their original composition, the residual deposits differ. Residuum from impure limestone, in which noncalcareous matter is an important constituent of the rock, is yellow-brown to red-brown clayey material rich in silt, sand, and gravel. Such residuum, in which there are few or no remnants of the original limy rock, occurs in very thick deposits distributed over wide areas. Residuum from pure limestone is fat red clay with no coarse particles. These clays occur as thin, discontinuous deposits, commonly associated with many exposures of the bedrock limestone.

The residual deposits of Okinawa are remarkably permeable in the natural state; where undisturbed, they drain and dry rapidly following rains. Surface drainage systems develop in regions underlain by thick deposits of sandy and gravelly residuum, but water flows in such systems only during and immediately after heavy rains. Areas underlain by limestone with a thin residual cover do not develop surface drainage systems; instead runoff disappears into sinks. In both types of regions, however, much meteoric water seeps through the residual cover into the underlying rock.

Descriptions of Rimming Walls

Walls Along Stream Banks

Description.—Rimming walls or ridges occur along the banks of many streams in Okinawa. They range from long continuous walls, easily traceable along both banks, to low, elongate hills along one bank only. All such hills are located on the edge of the bank, are elongate parallel to the course of the stream, and are found only in areas formerly covered with Pliocene or Pleistocene limestone. The most distinctive and best-developed examples of these walls occur along the banks of the Hiza-gawa and the Chatan-gawa. The locations of these and some of the other walls are shown in Figure 1.

Hiza-gawa.—The best and clearest examples of walls along a stream are those which flank the Hiza-gawa[1]. This stream is the largest

[1] Also known as the Bisha-gawa.

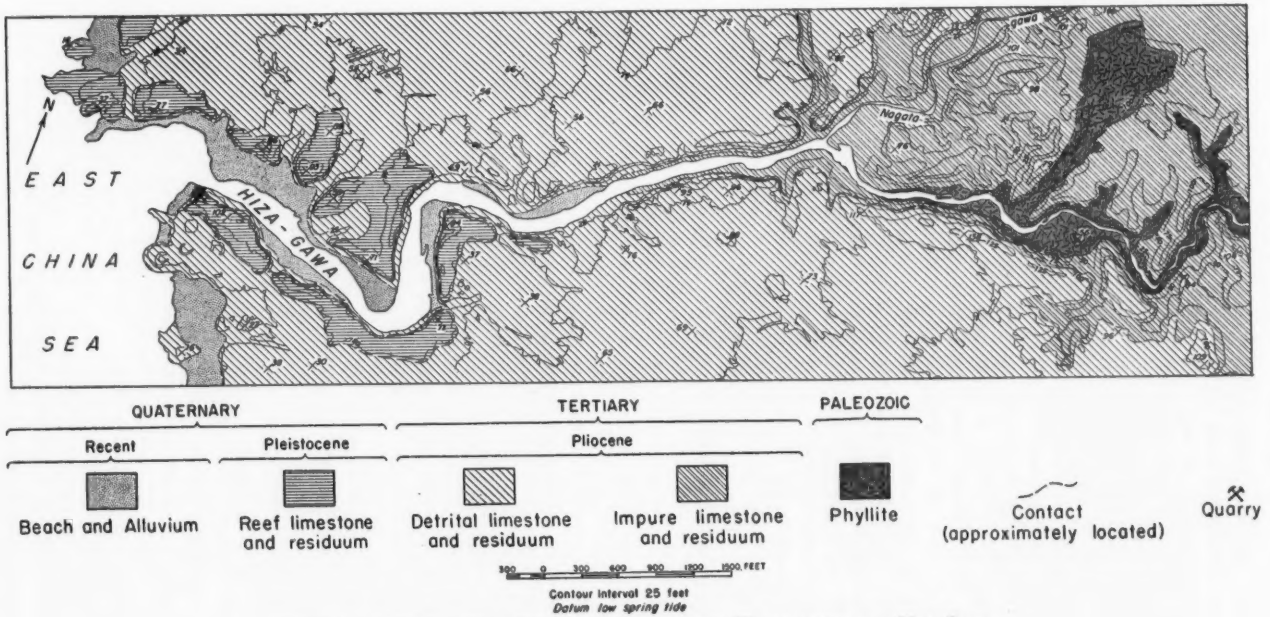

FIGURE 2.—GEOLOGIC AND TOPOGRAPHIC MAP OF THE WALLS ALONG THE HIZA-GAWA
See Plate 1.

stream in southern Okinawa, south of the isthmus of Ishikawa. It is formed in the center of the island by the junction of two small streams and flows westward, across the trend of the island. In its course, the Hiza-gawa follows an irregularly curving to sinuous route incised into the limestone terrace. A large tributary from the north, the Nagata-gawa, joins the stream about a mile above the mouth. From the mouth to a point a little upstream from the Nagata-gawa, the stream is tidal. In the tidal part, sandy beaches have been formed locally at the base of the steep to vertical canyon walls which rise from the water's edge. Above the tidal reaches, the stream gradient averages about 40 feet per mile. The canyon in the upper, nontidal part has a narrow V shape, with steep to vertical walls. Phyllite is exposed in the stream bed and locally in the lower canyon walls. The upper walls are cliffs of limestone or steep slopes of residuum from gravelly limestone.

Prominent ridges of limestone flank the Hiza-gawa along both the north and south banks. The ridges stand well above the surrounding country and follow closely the curves of the stream. The inner, streamward slopes of the ridges are steep to vertical; they are, in fact, the upward continuation of the slope of the canyon wall. The outer slopes, descending to the surrounding area, are less steep. The walls are more prominent and more continuous near the mouth of the stream, less so toward the head. The general effect of the stream and flanking ridges is that of a stream which follows a course cut along the center line of a curving, irregular line of hills (Pl. 1, fig. 1).

The wall along the north bank is generally less prominent than that along the south bank. The north wall is distinct and prominent only between the Nagata-gawa and the river mouth. In that area, it attains a maximum elevation of about 97 feet above stream level and stands 30 to 35 feet above the adjacent terrain. The wall varies in height but generally is about six times as wide as it is high. East of the Nagata-gawa the wall is indistinct or missing.

The south wall is well developed between the mouth and a point about 1 mile east of the Nagata-gawa. It has an irregular crest, but the elevation of the high spots of the wall rises from west to east. The relative elevation, however, is much greater in the west, where the wall is lower. Near the stream mouth, this wall stands 102 feet above the stream level and 90 feet above the neighboring land. Inland, just east of the Nagata-gawa, though this wall reaches an elevation of 135 feet above the stream level, it stands only 45 feet above the general level of the land to the south. This wall is, on the average, about four times as wide as it is high.

Both the north and south ridges are composed of similar limestones, though both include several different types (Fig. 2). The rock exposed includes highly impure to relatively pure, poorly bedded detrital, noncoralliferous Pliocene limestone and rubbly, poorly bedded Pleistocene limestone rich in coral fragments. The shape of the wall does not seem to be dependent on the type of limestone which forms it. Where the banks are composed of gravelly residuum from impure limestone, the ridges are either indistinct or absent. Though limestone formerly covered the entire area, very little is now exposed in areas away from the river banks. The surrounding country is covered by a deep mantle of red-brown clayey to gravelly residual soil.

A quarry in the west end of the south wall exposes a good cross section of the wall. The rock exposed in the quarry is rather pure, white, porous, rubbly, highly coralliferous limestone. The rock in the interior of the wall is poorly cemented and crumbly. Outward, toward the surface of the wall, cementation increases. The surface rock is hard, secondarily cemented, cavernous material, like all the other limestone exposed in the area.

Chatan-gawa.—Very prominent walls are developed along the lower courses of the

PLATE 1.—WALLS BORDERING STREAMS

FIGURE 1.—AERIAL PHOTOGRAPH OF THE MOUTH OF THE HIZA-GAWA
Showing the well-developed limestone walls along the stream.

FIGURE 2.—AERIAL PHOTOGRAPH SHOWING THE LIMESTONE WALLS AT THE MOUTH OF THE CHATAN-GAWA

FIGURE 1

FIGURE 2

WALLS BORDERING STREAMS

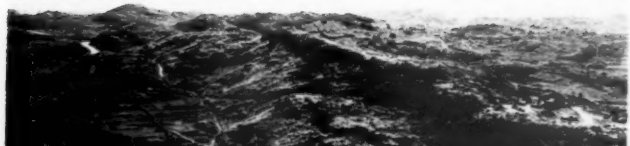

FIGURE 1

FIGURE 2

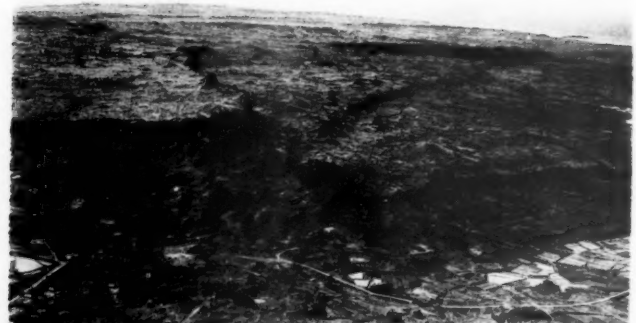

FIGURE 3

WALLS ALONG FAULT SCARPS AND AROUND A SINK HOLE

Chatan-gawa,[2] a small permanent stream about 2 miles south of the Hiza-gawa. This stream rises on the crest of the island and flows westward to the East China Sea. Though the stream, which drains an area of only 3½ square miles, is small, its flanking walls are even more prominent than those of the Hiza-gawa.

In its upper course, the Chatan-gawa flows southwest for about 1 mile and drops down a fault scarp with a steep gradient. At the foot of the scarp the stream makes a sharp bend and flows almost due west to the sea, a distance of about three-quarters of a mile. In the lower course, below the bend, the stream crosses a coastal flat, and the gradient is less than 40 feet per mile. The stream is tidal in the lower 900 feet.

Walls are well developed only along the lower course of the stream, where it crosses the coastal flat. There, the stream flows in an entrenched course in the floor of a wide, flat-floored canyon. The canyon walls are the inner faces of linear ridges of limestone that parallel the stream course and set it off from the coastal flats to the north and south. These ridges have steep inner slopes and generally less steep outer slopes. To the north, there is a broad, gently sloping, coastal terrace which reaches a maximum elevation of 25 feet above sea level. To the south, the 25-foot level is locally narrow but widens to give a wide area of coastal terrace. The low country on both sides of the stream makes the flanking ridges stand out in bold relief. The overall effect is that of a small stream crossing flat country with high levees along both banks (Pl. 1, fig. 2).

The floor of the canyon is filled to the 25-foot terrace level with alluvial and coastal flat deposits into which the stream course has been incised 5 to 6 feet. When heavy rains fall, the stream rises, overflows its banks, and floods most of the canyon floor.

[2] Named Shirahi-gawa on some maps.

Near the mouth of the stream, the south wall reaches an elevation of 140 feet above the stream flat and is about 450 feet wide at the base. This wall may be traced inland with only minor breaks until it becomes indistinct in the dissected country along the fault scarp. The north wall rises only 65 feet above the stream flat at the western end. To the east, it rises to match and surpass the elevation of the south wall but is not so distinct and clear-cut a feature. It ends against the fault scarp. The north wall is unusual at its western end, in that the inner and outer slopes are equally steep. This is believed to be due to marine abrasion of the outer side of the wall at the time the 25-foot terrace was cut.

Both the north and south walls are composed of impure, porous, silty and sandy, cross-bedded, detritial limestone with much terrestrially derived debris intermixed with the marine calcareous material. The upper part of the south wall, however, is composed of purer, silty limestone with only minor amounts of debris from the land. Similar silty limestones in the same relationship occur in the surrounding region and evidently once covered the entire area. The purer limestone is now limited to the south wall because it has been stripped from the northern wall.

All the rock exposed is secondarily cemented, but the lower, more impure limestone is neither as well nor as uniformly cemented as the overlying purer rock.

Walls along Fault Scarps

Description.—Southern Okinawa is broken into a number of tilted fault blocks by steeply dipping normal faults. Many of the fault scarps are topped with linear, parapetlike ridges of pinnacled limestone that stand like serrated knife edges above the general level of the upthrown block. A noteworthy feature of some of these fault-scarp ridges is that the

PLATE 2.—WALLS ALONG FAULT SCARPS AND AROUND A SINK HOLE

FIGURE 1.—AERIAL PHOTOGRAPH LOOKING SOUTHEAST ALONG THE FAULT SCARP AT MACHINATO
Note the ridge along the crest of the scarp.

FIGURE 2.—AERIAL PHOTOGRAPH OF YAEJU-DAKE (left center) AND YUZA-DAKE (right)
Note the linear ridges along the scarps.

FIGURE 3.–SUBCIRCULAR WALL AROUND THE SINK HOLE NEAR KAWASAKI
The quarry in the foreground exposes limestone in the center of the wall.

limestone is preserved on the upthrown side and is removed on the downthrown side. This is the opposite of what might be expected in a region as humid as Okinawa and is the antithesis of the familiar textbook illustration of the erosion of limestone in fault blocks, where limestone is preserved in the downthrown block and removed from the upthrown. Fault-scarp ridges are developed in many places in southern Okinawa. The locations of some of the well-developed ridges are shown on the map (Fig. 1).

Escarpment near Machinato.—A well-defined ridge caps the prominent fault scarp that reaches the west coast at Machinato. This scarp is 150–175 feet high, faces northeast, and trends north-northwest. It is continuous for 2½ miles from the west coast inland to its abrupt termination near the crest of the island. The ridge along the top of the scarp is a narrow line of elongate hills that stands well above the terrain immediately to the south. The ridge averages about 50 feet and reaches 100 feet higher than the adjacent terrain on the south. The north slope of the ridge is continuous with the scarp face and generally is a little steeper than the south slope of the ridge, though both are steep. The ridge is, on the average, about four times as wide as it is high, but the crest of the ridge, composed of jagged pinnacled limestone, is seldom more than twice as wide as it is high (Fig. 3; Pl. 2, fig. 1).

Secondarily cemented Pliocene limestone makes up most of the ridge. In the southeast the Pliocene limestone is only a cap above a ridge of Miocene sediments, a ridge which owes its relief to the resistance of the limestone to erosion. Pleistocene limestones overlie the Pliocene in the northwest. The dissected tilted block to the south is underlain by Miocene silty sand and gray clay with a local capping of limestone or residual silty clay. It is clear that the block was formerly covered by a continuous sheet of limestone which has now largely been removed. The downthrown block is capped by limestone under an overburden of residuum.

Yuza-dake and Yaeju-dake.—Two peaks, Yuza-dake 548 feet) and Yaeju-dake about 595 feet), are the highest points of the upland area of southernmost Okinawa. Both are high points on well-developed linear ridges that cap north-facing fault scarps. The ridges are long, narrow, wall-like features that stand as much as 50 feet above the general surface of the upthrown block and are almost as high as they are wide at the base. They are continuous along the fault scarps for about a mile (Pl. 2, fig. 2).

The ridges are jagged, pinnacled masses of secondarily cemented lagoonal and reef limestone. The limestone forming the top of the walls is Pleistocene reef limestone composed of large masses of coral, in position of growth, surrounded by rubbly detritus of reef origin. The detrital rubble includes algal, molluscan, and foraminiferal material in addition to coral fragments. Underlying the reef limestone is a finer-grained, detrital limestone of Pliocene age that contains algal, foraminiferal, and molluscan matter but no coral; all the limestones in this area are pure.

The fault blocks immediately south of the ridge have little relief. Residuum forms only a thin, discontinuous layer over these surfaces, probably because the limestone was pure. Farther south a topography of limestone ridges and valleys has been developed with coral limestone exposed in only a few of the ridges. Evidently the entire block was once covered by a layer of coral-reef limestone overlying non-coralliferous limestone and the present surface has formed by erosion.

Subcircular Wall Rimming a Sink

Perhaps the most striking example of a rimming wall on Okinawa is the subcircular wall near the village of Kawasaki. There a sink hole occupies the center of a small conical hill, like a crater in a cinder cone (Fig. 4; Pl. 2).

The hill containing the sink is one of several prominent hills of horizontally layered, detrital limestone that jut above the residuum-covered terrain as tree-covered knobs, remnants of a continuous layer. The hill is about 700 feet in diameter at the base and stands 80 to 90 feet above the surrounding land with slopes that average about 30°. The sink hole is located centrally in the hill mass, offset slightly to the northwest. The hole itself is typical of many on the island, differing only in its unique position. It measures about 300 feet in diameter, and its bottom is at about the same elevation as the surrounding country. The hole has steep

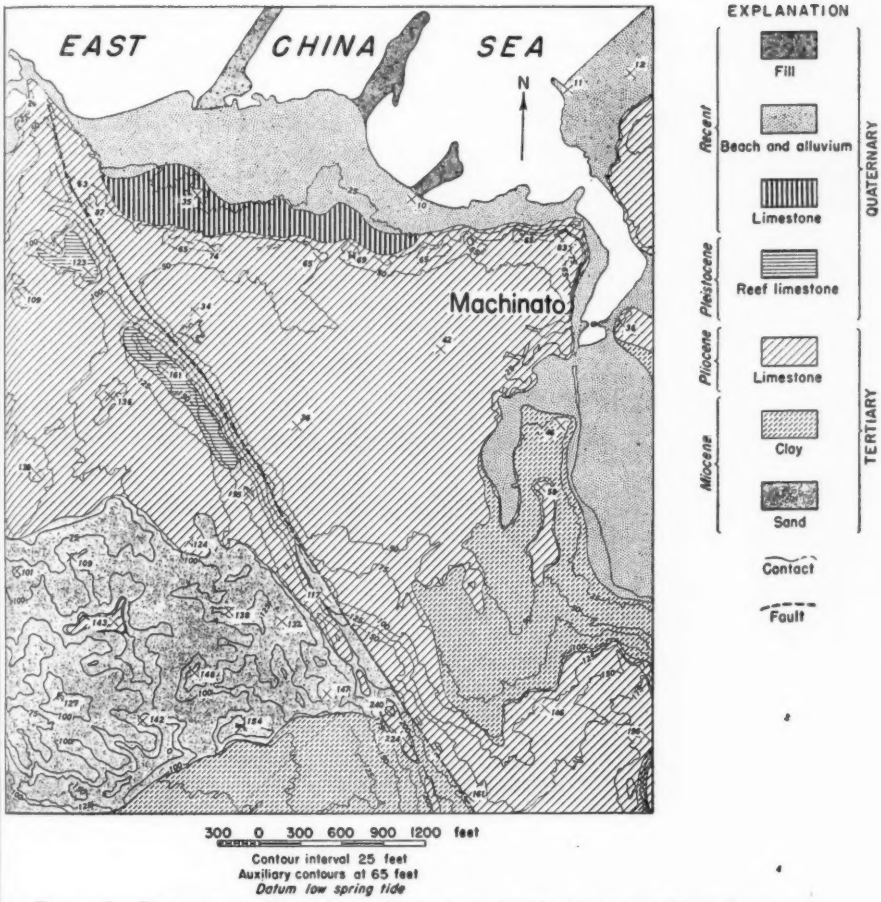

FIGURE 3.—GEOLOGIC AND TOPOGRAPHIC MAP OF THE WALLS ALONG THE FAULT SCARP AND TERRACE EDGE NEAR MACHINATO

See Plate 2.

to vertical sides of hard, well-cemented limestone, locally covered with patches of dripstone. The floor is covered with residuum and rubble from the construction of a road into the floor of the hole.

In cross section the hill has the characteristics of the walls along the river banks. It is steep on the inner face, less steep on the outer, composed of limestone, and is marginal to a drainage feature.

A quarry in the northwest flank of the hill exposes the rock which forms the wall. Very impure limestone that contains many pebbles and cobbles of Paleozoic rock is exposed at the base of the quarry face. Overlying that is a layer of poorly bedded detrital limestone. Scattered sand grains and pebbles occur near the base of the detrital layer, but the bulk of the layer contains little noncalcareous matter. The top of the quarry face, as well as the top of the hill, is rubbly, limestone containing many algal-nodules. The limestone in the

interior of the wall is less well cemented than the rock exposed at the surface, but all the rock is partly cemented.

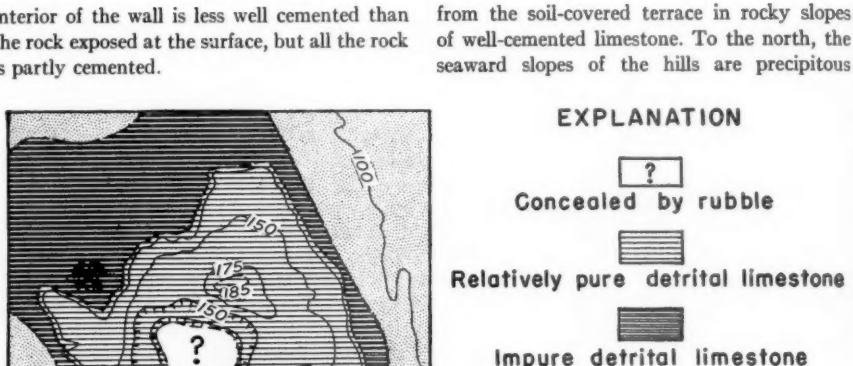

FIGURE 4.—GEOLOGIC AND TOPOGRAPHIC MAP OF THE WALL AROUND A SINK HOLE NEAR KAWASAKI
See Plate 2.

Rimming Walls Along Terrace Edges

General statement.—Lines of low hills occur along the outer edges of several terraces on Okinawa and are believed to be equivalent to the rampart walls of other Pacific Islands. These ridges or lines of elongate hills range in elevation from a few feet to more than 40 feet above the contiguous terrace surface. Two of the better-developed examples are described.

North coast of the Motobu Peninsula.—Pliocene and Pleistocene limestones form a broad terrace along the north side of the Motobu Peninsula. The terrace has a gently irregular surface, most of which is covered by residual soil. North of the town of Yonamine, the outer edge of the terrace is marked by a line of hills that rise about 40 feet above the general terrace level. The hills rise sharply from the soil-covered terrace in rocky slopes of well-cemented limestone. To the north, the seaward slopes of the hills are precipitous cliffs that drop down to a narrow, lower terrace level, about 20 feet below the broad surface. The lower level has a veneer of residual soil through which project ledges of limestone. This lower terrace level is found only locally along the coast; elsewhere it has been destroyed by later marine abrasion.

It is interesting to note that a ridge is also present along the east bank of the Shigemagawa, a north-flowing stream 1 mile west of Yonamine. The ridge rises to about the same elevation as the hills along the terrace edge and like those hills is composed of cemented detrital limestone. The summit level of the two ridges probably approximates the original level of the limestone terrace before subaerial erosion.

Machinato.—A line of low hills marks the break in slope between two terrace levels in

the embayment north of the escarpment at Machinato. The upper level is about 40 to 50 feet above sea level, the lower level 25 to 35 feet above sea level. The hills that occur along the break in slope reach elevations of more than 80 feet above sea level (Fig. 3).

These hills unlike those near Yonamine do not have steep seaward slopes, except at the north end of the ridge. This appears to be due to late Quaternary deposits that accumulated against the earlier cliffs and smoothed the abrupt break in slope. At the north end of the ridge, recent marine erosion removed the Quaternary deposits and the 25- to 35-foot terrace and cut a cliff that rises from the recent beach level to the hillcrest.

Summary

All these rimming walls have certain characteristics in common:

(1) Walls occur in porous, permeable detrital limestones, which were deposited as poorly cemented aggregates of calcareous materials and which cement themselves upon exposure to alternate wetting and drying. The walls are composed of limestone that is well cemented on the surface but, where exposed in excavations, is less well cemented in the interior.

(2) Walls are erosional remnants of formerly continuous layers of similar limestone. The walls rise above areas of residuum from limestone, and, where exposures of limestone occur, the rock is similar to that in the walls.

(3) There is no original sedimentary control of these features. Walls occur in all types of Pliocene and Pleistocene limestone, and some walls include several types. The position and shape of the walls is not controlled by reefs or any special conditions of sedimentation.

(4) Walls occur where a steep face of limestone was exposed; that is, in canyons, along fault scarps, in a sink hole, and along the cliffs which mark the seaward edge of terraces.

Theory of Origin

Rimming walls are believed to form by surface-controlled secondary cementation of an original steep slope followed by differential erosion which brings the cemented, denser zone into relief. This process is believed to occur in the normal weathering cycle of porous, poorly cemented, fragmental limestones under tropical or subtropical conditions.

Both the lagoonal and reef limestones were deposited as soft, unlithified, porous, permeable sediments and have not been consolidated by regional diastrophic processes. The rock, however, cements itself along any surface exposed to alternate wetting and drying. This surface-controlled cementation or casehardening is believed to occur in the following manner. Rain water containing carbon dioxide falls on the rock, dissolving calcium carbonate as it runs through the pores. After the rain stops, any surface exposed to the air becomes a drying surface from which water evaporates. As the water evaporates, carbonate, held in solution as calcium bicarbonate, is precipitated as calcite. The precipitated calcite is concentrated in the tiny cracks between grains and over the surface of the individual particles. This ties the particles together and cements the rock. The result is a highly permeable but hard limestone. Aragonite in the original sediment goes into solution more readily than calcite and is deposited from solution as calcite. This recrystallization greatly aids the process of secondary cementation. In cut blocks, where no lime is added, the rock may be cemented with no increase, or even a loss, of density. In natural exposures, where additional water reaches the face by either capillary action or gravity, the density of the rock is generally increased in the cemented zone.

Erosion through solution by rain and ground water takes place intermittently with the secondary cementation. Much of the lime dissolved in the percolating rain water is carried off, leaving only the insoluble material as a residuum. On Okinawa, the residuum from limestone is itself very permeable, and much of the rain water that falls on it sinks through into the underlying bedrock. As a result, the residuum is not washed away but collects, first in pockets, later as a continuous mantle, on all level and nearly level surfaces. The layer of residual material holds moisture in contact with the underlying rock and greatly inhibits the process of secondary cementation. Vegetation soon takes hold in the residual soil, and organic waste accumulates. Respiration of plant roots and the decomposition of organic

material raises the concentration (partial pressure) of carbon dioxide in the soil air. The partial pressure of carbon dioxide in the gaseous phase in equilibrium with water determines the amount of carbon dioxide that will be dissolved by the water. Therefore rain water, saturated with carbon dioxide under atmospheric conditions, will dissolve additional carbon dioxide as it moves through the soil where the concentration of carbon dioxide is higher than it is in the air. As the solubility of calcium carbonate in water depends on the amount of dissolved carbon dioxide, rain water that has passed through soil can dissolve much more calcium carbonate than rain water that has been in contact only with atmospheric air (Foster, 1950, p. 36). This results in more rapid solution of limestone in areas under a soil cover. The combination of increased solution and decreased cementation under soil cover results in accelerated erosion of such areas.

Soil does not accumulate on steeply sloping faces, however. There, what residue is formed is washed away, and the exposed face is kept free to act as a drying surface. As a result, secondary cementation continues to be the dominant effect though solution is still active on the surface. Continued cementation results in a hard zone of considerable thickness which grades inward to uncemented rock. Secondarily hardened zones are more dense and therefore less rapidly removed by solution than the uncemented rock. Steeply sloping hardened zones also lack soil and therefore are subject to less solution than the surrounding soil-covered areas.

The two processes, the hardening of steep faces and the increased rate of solution in the surrounding, flatter areas, result in differential erosion which brings the harder areas into relief. As the rim area comes into relief, the reverse surface is exposed to secondary cementation, and the resulting hill develops a complete carapace of harder, denser material.

All outcrops of limestone are subject to solution by rain, and the secondarily cemented walls are no exception. The same surface that is being cemented during dry spells is dissolved during rains. As a result the cemented rock of the walls exhibits jagged solution surfaces. This gives rise to the seeming paradox that the best-cemented rock shows evidence of intense solution.

This process explains the location, lithology, and structure of the rimming walls of Okinawa. It also explains why limestone ridges remain along the upthrown sides of some faults while all the limestone is removed from the downthrown side. A limestone ridge is formed along the fault scarp by secondary cementation of the exposed limestone face, followed by differential erosion. The dense, cemented limestone of the ridge is subject to solution by rain, aided only by a sparse growth of plants. The limestone of the downthrown block, however, is soon covered by soil which accelerates the rate of solution. If the water table is low enough to permit the acid water from the soil to reach the limestone, all the lime will be dissolved from the downthrown block long before the rain dissolves the ridge along the scarp.

Other Examples of Limestone Walls

Rampart Walls of the Pacific

General discussion.—It is believed that the process of secondary cementation and differential erosion occurs in the normal course of erosion of all porous, permeable unlithified limestone if the climate is such that the rocks are intermittently wet and dried. Tropical and subtropical climates with heavy showers of short duration followed by periods in which the sun bakes the ground are most favorable. The existence of walls along the outer margins of terraces on widely spaced "high limestone islands" in the Pacific bears out this belief. The process has a wide geographic range as evidenced by the following examples.

Kita Daito Island.—Kita Daito is an elevated atoll about 200 miles east of Okinawa. The atoll rim has been cut into a number of terraces by marine erosion during stands of the sea below that reflected in the uppermost level or terrace of the atoll. Spectacular walls are developed along the inner and outer edges of the upper terrace surface, and smaller walls have been formed on some of the lower surfaces. Earlier writers described these walls as emerged coral reefs, but Saplis and Flint (1949) showed them to be independent of reef structures and

attributed their formation to differential solution. They postulated surface-controlled secondary cementation as the controlling factor in the location and development of these walls. This theory was strengthened by finding a Japanese tunnel in one of the walls that showed the rock of the interior to be less well cemented than the rock on the surface.

Angaur Island, Palau Group.—The topography and geology of Angaur in the Palaus have recently been described by Irving (1950). He points out the presence of a series of concentric ridges and intervening basins which ring the highest portion of the island. The ridges are highest in the center and decrease in elevation progressively outward. In describing these features, Irving says (p. 26):

"Although variations in the degree of cementation of rock of the limestone platform may conceivably be explained in part by the operation of secondary processes, it seems doubtful that all the described relations can be thus explained. The concentric ridges, which are notably composed of compact limestones, are predominantly areas which at the same time display the effects of intense solution. The basin areas, on the other hand, should be comparatively more favorable loci for deposition of secondary calcium carbonate. Nevertheless, it is around the basin areas where the less-consolidated limestones are more likely to be found."

Irving further states (p. 36–37) that the inner ridge is the best-cemented, the outer ridges progressively less well cemented, and the rock under the flats is least cemented of all. Fossils are least well preserved in the innermost ridge and better preserved in the outermost.

All these facts are consistent with the theory of surface-controlled secondary cementation and differential solution. The innermost ridge represents the highest and therefore the oldest terrace wall. Because it is the oldest it is the best-cemented ridge and since it has been exposed the longest it is the ridge most deeply attacked by rain-water solution. The outer ridges, representing progressively more recent sea cliffs, are both less well cemented and less attacked by rain. The basinal areas and the flats have the deepest cover of soil and are least cemented.

Tuvuthá and Wangava, Fiji.—Prominent rampart walls have been described from the islands of Tuvuthá and Wangava by Hoffmeister and Ladd (1945), who recognized that the walls were formed by differential erosion of the areas behind the sea cliffs but offered no explanation for that localization of solution. Surface-controlled secondary cementation along the sea cliffs can explain the greater resistance of the cliffs to solution and their weathering into relief.

Eua, Tonga.—Hoffmeister and Ladd described prominent rampart walls along the terrace edges of the eastern coast of the island of Eua in the Tonga group (1945). They attribute these walls to differential solution and very plausibly explain the formation of the walls as due to accelerated solution of the down-dip portions of a tilted block. The process is, they believed, analogous to the solution by rain of sloping faces of impermeable limestone. An experiment proved the effectiveness of such a process to produce a rim along the upper edge of a sloping face of lithographic limestone, if the surface has the proper inclination. Displaced terrace levels on Eua indicate a tilt of the island down to the west, which conforms with the theory. They then conclude (p. 87).

"the existence of a limestone wall on the outer edge of a terrace may be a significant factor in determining differential uplift of an island. In other words if an island possesses a terrace profile similar to that of Eua, then it may well be indicative of island uplift. It is believed that marginal walls may form on terrace edges without island tilting during uplift. In this case, however, the terrace profile is exemplified by the terraces of Tuvuthá and Wangava".

This conclusion should be viewed with caution insofar as taking the presence of terrace walls as evidence of tilting. The walls of Eua appear to have much the same shape as those on Kita Daito and Tuvuthá, where there has been no tilting. Tilting, therefore, is not necessary for their formation, as was pointed out by Hoffmeister and Ladd. It follows then, that the presence of walls cannot be taken as evidence of tilting. Of course, tilted terrace levels can be taken as evidence of tilting, and the latter part of the conclusion is true.

It seems probable, to the writers that the rampart walls of Eua were formed by the same processes as those of the other islands of the Pacific and that the tilting was coincidental.

Rimming Walls of Tertiary Age

Surface-controlled secondary cementation is effective in a climate where hot sunshine alternates with drenching showers. The converse should also be true, though it is thus far not proved: the presence of surface-controlled secondarily cemented limestone in the stratigraphic column may suggest the nature of the paleoclimate. Differential erosion of limestone, however, may occur under a great range of climatic conditions and gives little evidence of climate. It is quite possible for limestone cemented under one set of climatic conditions, to be differentially eroded and brought into prominence under a later, different set. Just such a case has been reported by Burnaby (1950) from the north coast of Norfolk, England.

Near the village of Sheringham, Norfolk, a subaerial erosion surface was developed on the Chalk formation, before the deposition of the overlying Weybourne Crag formation. Small tubular solution pipes, open at the top, were formed extending a few yards into the chalk. The walls and bottoms of these pipes were converted into hard tubular structures by secondary lithification which also formed a thin layer along the upper surface of the chalk, between the pipes. The whole was buried by the succeeding deposits, the basal conglomerate of which filled the open tubes. Erosion is now attacking cliffs of the chalk and the overlying Weybourne Crag formation. The harder denser pipes of cemented chalk weather out from the uncemented chalk and stand above the ground surface as tubular, gravel-filled, limestone stacks.

Burnaby reviewed the possible causes of cementation in chalk and decided that the lithification was probably due to the action of organic acids. He further concluded that it gave a clue to the climate of the pre-Weybourne Crag erosion period.

The writers agree with the conclusion but believe surface-controlled cementation due to alternate wetting and drying best explains the lithification of both the pipes and the intervening surface.

Conclusion

The rimming limestone walls of Okinawa and of other Pacific islands have formed by surface-controlled secondary cementation and differential erosion. Surface-controlled cementation occurs along steep exposures of poorly consolidated limestone which is exposed to alternate wetting and drying. Secondary cementation is more rapid in those limestones with many fine particles, those containing aragonite, and those relatively free from noncalcareous debris. Differential erosion of the cemented zones is due in part to their greater density and in part to accelerated erosion in the soil-covered surrounding areas.

All exposures of limestone are subject to solution by rain water and develop highly fretted surfaces. Such evidence of solution on the limestone walls obscures the effects of cementation though cementation is the dominant process.

The presence of local cemented zones in poorly consolidated limestones suggests exposure to alternate wetting and drying. Well-developed surface-controlled secondary cementation may be indicative of a tropical or subtropical climate.

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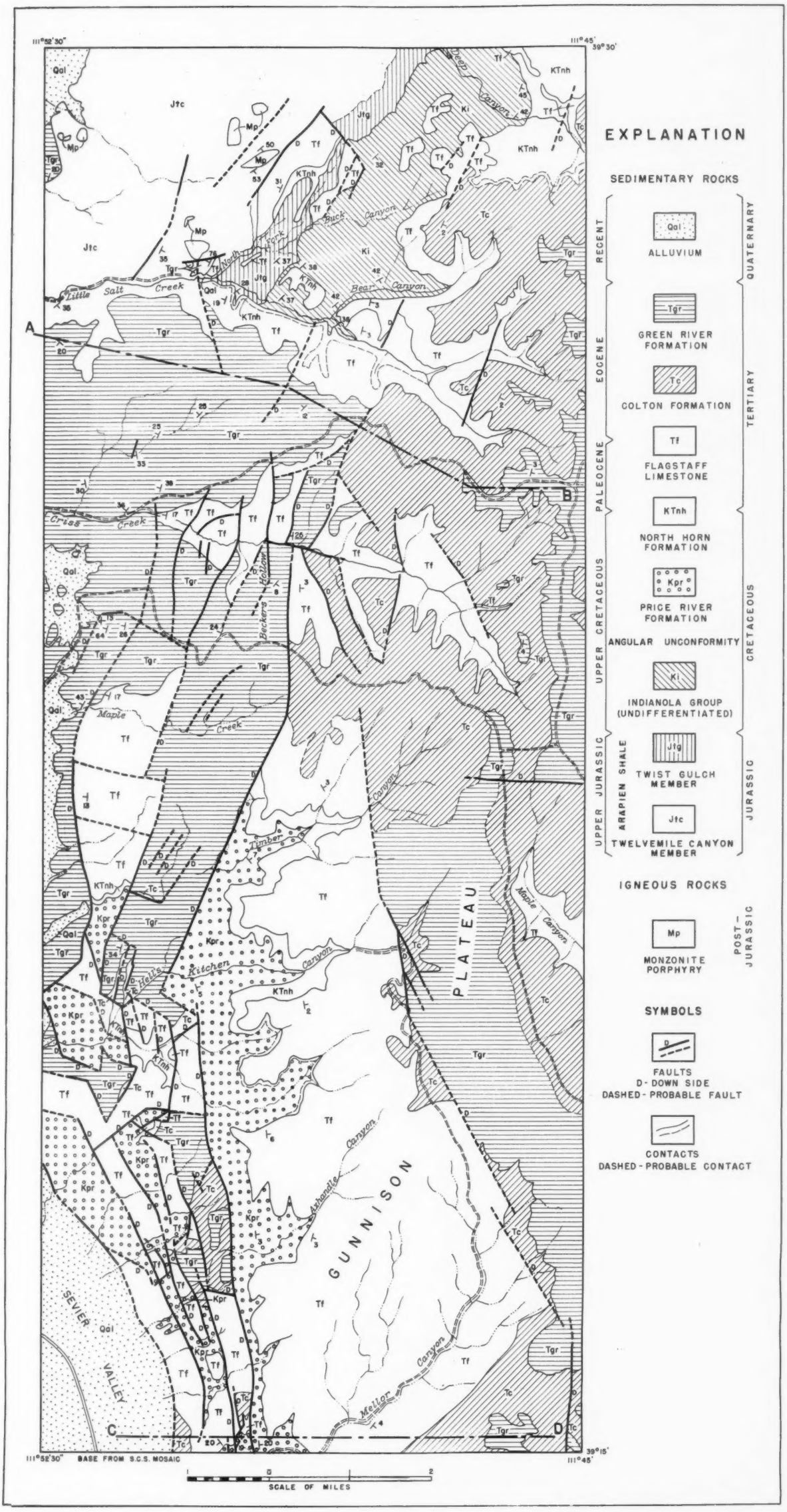
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Tc COLTON FORMATION
Tf FLAGSTAFF LIMESTONE
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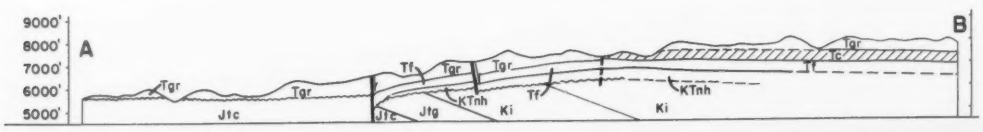
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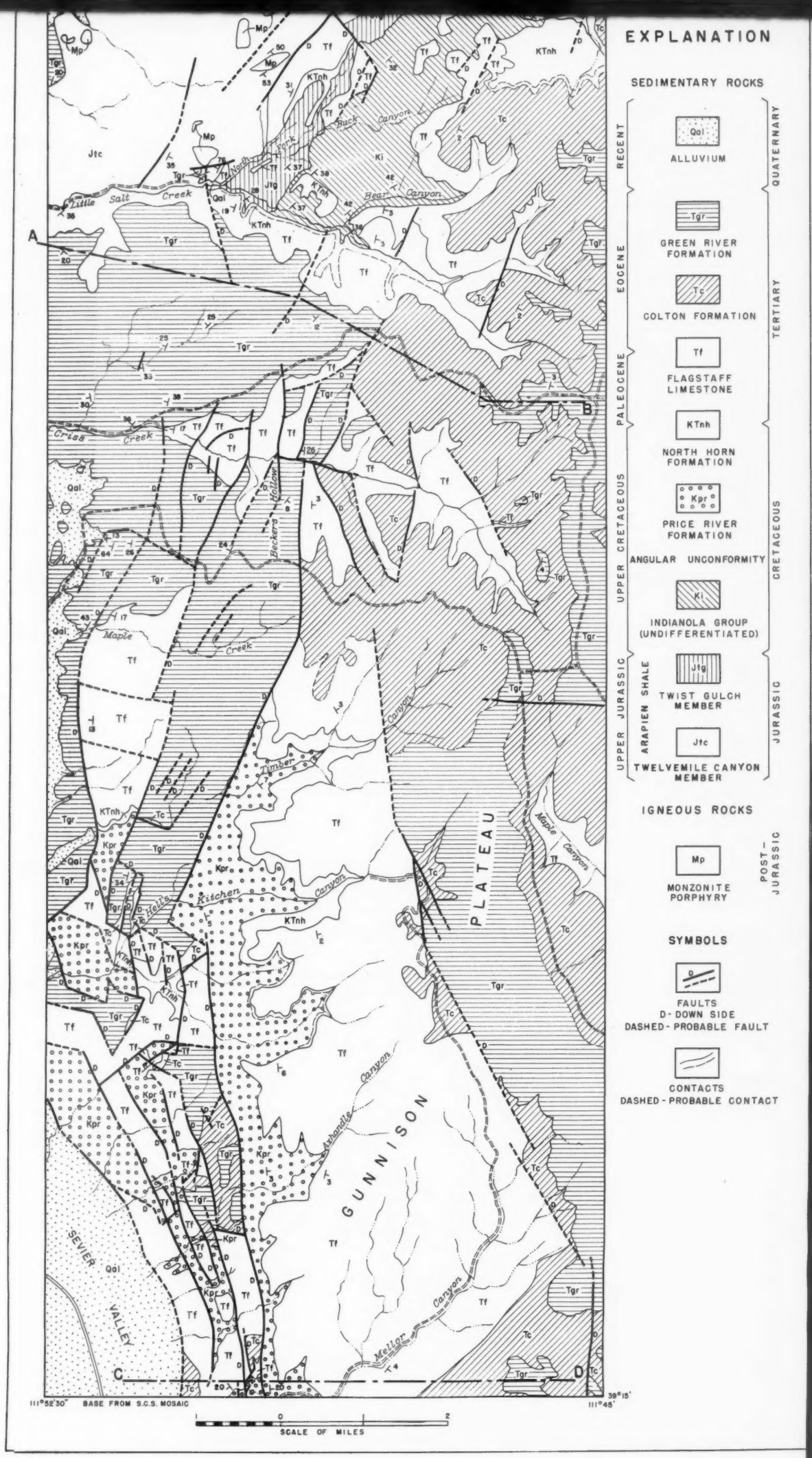
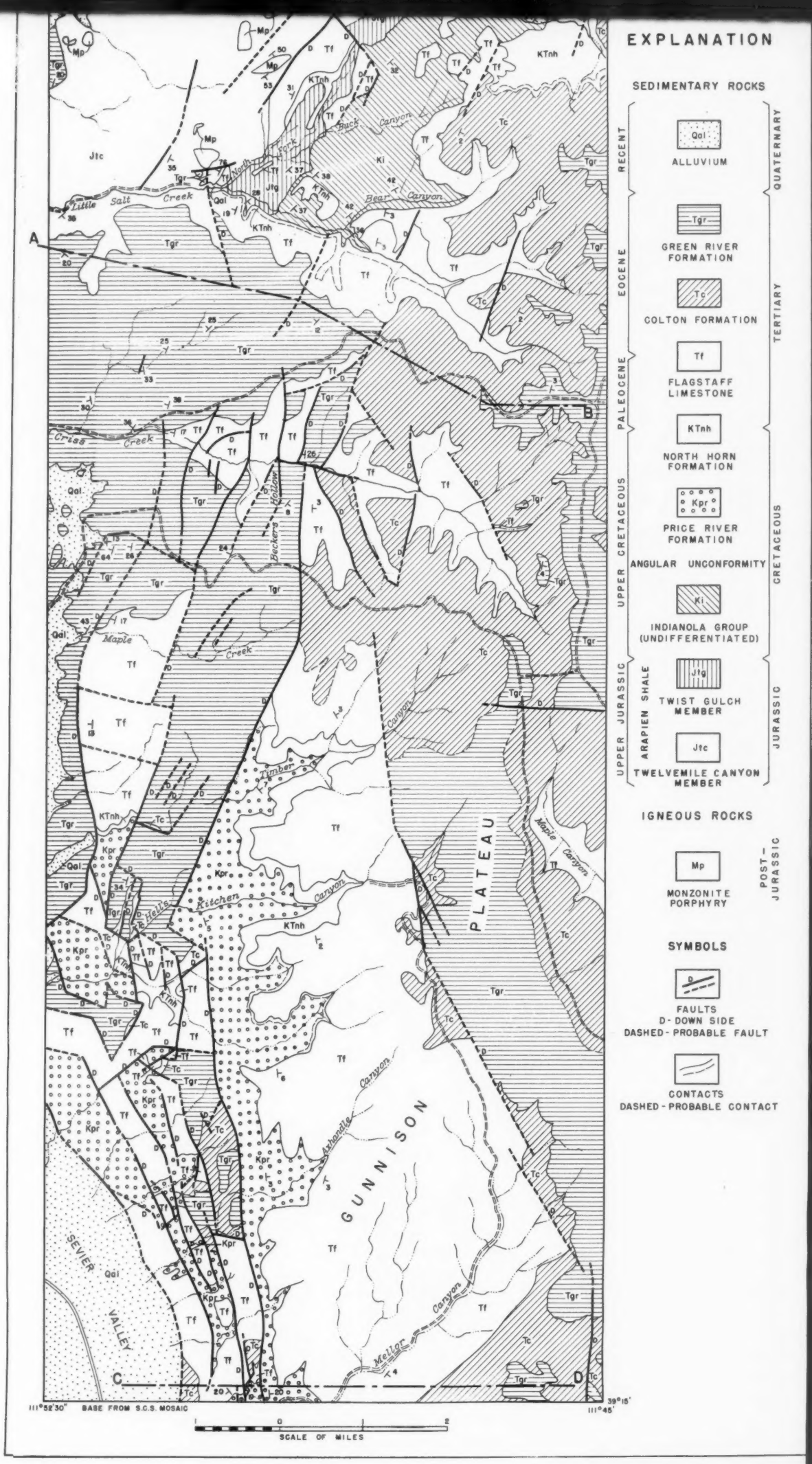
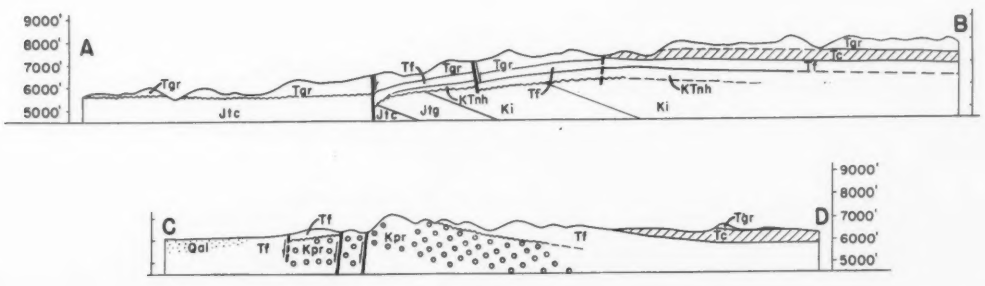


GEOLOGIC MAP OF THE WEST-CENTRAL PART OF THE GUNNISON PLATEAU, UTAH





GEOLOGIC MAP OF THE WEST-CENTRAL PART OF THE GUNNISON PLATEAU, UTAH

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GEOLOGY OF THE WEST-CENTRAL PART OF THE GUNNISON PLATEAU, UTAH

BY CLYDE T. HARDY AND HOWARD D. ZELLER

ABSTRACT

A detailed study of the west-central part of the Gunnison Plateau, Utah, has disclosed stratigraphic and structural relations important in the geological history of central Utah. The area mapped includes the eastern half of the Axtell No. 2 quadrangle, Manti area (U. S. Dept. of Agriculture, Soil Conservation Service).

The bedrock ranges in age from the Arapien shale (Upper Jurassic) to the Green River formation (Eocene). The North Horn formation (Cretaceous-Tertiary), the Flagstaff limestone (Paleocene-Eocene), and the Green River formation (Eocene) successively overlap the Arapien shale and the Indianola group in the northern part of the area. In the southern part of the area, the North Horn formation and the Flagstaff limestone successively overlap the Price River conglomerate with angular discordance. The latter relation establishes a post-Price River and pre-North Horn orogenic movement in central Utah.

A conspicuous west-dipping monocline, broken by many high-angle faults and a graben, extends along most of the western margin of the area. This structure is similar to the monocline and graben in the Wasatch Plateau to the east. At one point in the northern part of the area, the North Horn and the overlying Flagstaff oppose the Arapien shale across a high-angle fault. The Green River formation extends across the fault and overlaps the Arapien. This relation suggests faulting between Flagstaff and Green River time because the Colton formation, which normally occurs between the Flagstaff and Green River, cannot be differentiated in this immediate area.

Numerous small intrusive masses of monzonite porphyry of post-Upper Jurassic age occur in the Arapien shale.

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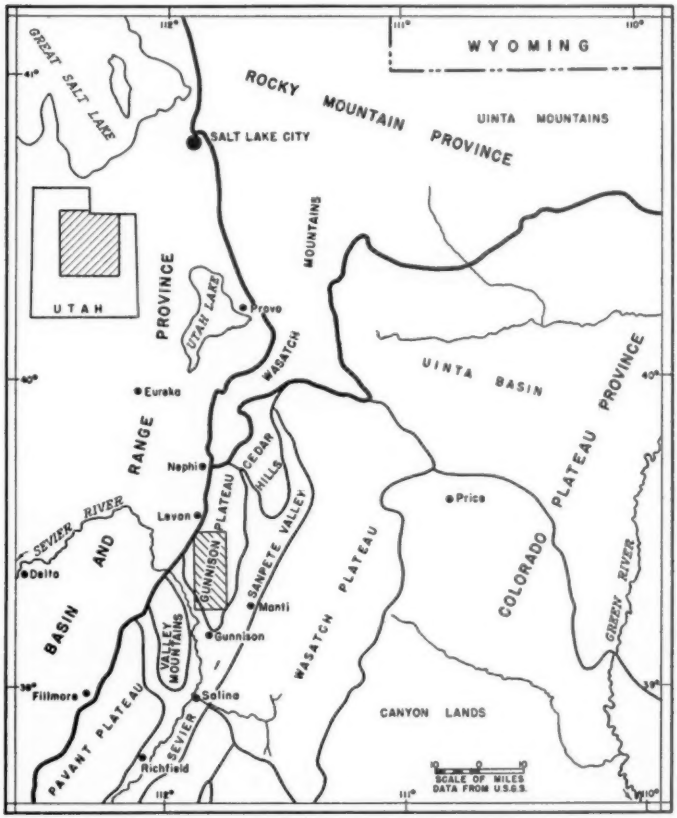

FIGURE 1.—INDEX MAP OF CENTRAL UTAH

INTRODUCTION

In 1947 and 1948 the writers independently investigated the stratigraphy and structure of adjacent areas in the west-central part of the Gunnison Plateau, Utah. These studies are summarized in this paper.

LOCATION

The Gunnison Plateau, about 100 miles south of Salt Lake City, extends southward from near Nephi, Utah, for 35 miles to Gunnison, Utah, and is about 11 miles in width. The plateau is separated from Mt. Nebo, at the southern extremity of the Wasatch Mountains, by Salt Creek Canyon, east of Nephi, Utah.

The area described in this report (Fig. 1) extends from near Levan, Utah, southward to Fayette, Utah, which is about 5 miles northwest of Gunnison. It extends eastward to the divide in the central part of the Gunnison Plateau. The area includes the eastern half of the Axtell No. 2 quadrangle, Manti area (U. S. Dept. of Agriculture, Soil Conservation Service), and is about 6 miles wide and 16 miles long.

MAJOR GEOLOGIC FEATURES

The exposed bedrock (Table 1; Pl. 1) ranges in age from Upper Jurassic (Arapien shale) to Eocene (Green River formation). The Arapien shale forms rugged foothills in the northwestern part of the area and contains numerous small

intrusive masses of monzonite porphyry. The Indianola group (undifferentiated) of Upper Cretaceous age overlies the Twist Gulch member of the Arapien shale in apparent conformity. The Indianola conglomerate forms an imposing cliff in the northwestern part of the plateau and in general dips gently eastward.

In the northern part of the area the North Horn formation (Cretaceous-Tertiary), the Flagstaff limestone (Paleocene-Eocene), and the Green River formation (Eocene) overlap the Arapien shale successively with distinct angular discordance. The Indianola group is also overlain in angular relation by the Flagstaff limestone and the North Horn formation. In the southern part of the area a conglomerate, which presumably represents the Price River formation, forms a high west-facing escarpment. The Price River is overlain by the Flagstaff limestone, in distinct angular discordance, and likewise by the North Horn formation a few miles to the north. Considerable relief on the post-Indianola and post-Price River erosion surfaces has influenced the distribution of later formations. The Colton and Green River formations overlie the Flagstaff limestone chiefly in the eastern part of the area.

A broad anticlinal structure trends northward, and the east limb forms the west flank of a syncline which occupies the central part of the Gunnison Plateau. The west flank of this anticlinal structure is broken by many high-angle normal faults and at least one

TABLE 1.—FORMATIONS IN WEST-CENTRAL PART OF THE GUNNISON PLATEAU, UTAH

System	Series		Unit	Lithologic characteristics	Thickness (Feet)
Tertiary	Eocene		Green River fm.	3. Tawny limestone, some tawny sandstone 2. Green-gray shale, tawny sandstone 1000 1. Light-gray argillaceous thin-bedded limestone, green-gray shale, gray sandstone 400	1400+
			Colton fm.	Red and green-gray shale, gray and green sandstone, gray limestone	0–600
	Paleocene		Flagstaff limestone	Gray limestone, yellow and red arenaceous limestone, some gray sandstone, dense gray limestone	30–750
			North Horn fm.	Gray and red sandstone, conglomerate, red shale, gray limestone	0–400
Cretaceous	Upper Cretaceous		Price River fm.	Conglomerate, gray and brown sandstone	1500+
			Indianola group	Conglomerate	5200+
Jurassic	Upper Jurassic	Arapien shale	Twist Gulch member	Reddish-gray sandstone, red siltstone	1900
			Twelvemile Canyon member	3. Gray shale and thin-bedded sandstone 2. Gray and red shale with gypsum 1. Dark-gray limestone	2500+

conspicuous graben with displacement of about 1000–1500 feet. Along the western margin of the central part of the Gunnison Plateau, the Flagstaff limestone and related formations dip westward in a pronounced monoclinal flexure.

Topography and Drainage

The Gunnison Plateau is a relatively narrow upland which rises from the Sevier Valley near Gunnison, Utah, to an elevation of almost 10,000 feet near the northern end. The surface of the plateau is maturely dissected. The margins are, in general, steep although lower foothills parallel the conglomerate escarpments along the western side. Elevations range from about 5200 feet in Sevier and Juab valleys, west of the plateau, to about 8900 feet in the central portion.

The west-central part of the Gunnison Plateau is dissected by eight major canyons which trend nearly at right angles to the margin of the plateau. Small streams in the five northern canyons are fed by springs and flow at diminishing rates throughout the summer. This area lies almost entirely west of the drainage divide in the central part of the Gunnison Plateau.

Previous Investigations

E. E. Howell prepared several excellent cross sections of the Gunnison Plateau, which were included by Dutton (1880, pl. facing p. 162) in the *Report on the geology of the High Plateaus of Utah*. Meinzer (1911, p. 67–74) briefly described the western margin of the Gunnison Plateau in a report pertaining to water resources of Juab, Millard, and Iron counties, Utah. Recent work in the Gunnison Plateau was initiated by Gilliland in 1946 as part of a survey of the Gunnison quadrangle (Gilliland, 1951). R. E. Hunt (unpublished dissertation) completed a study of the northern part of the Gunnison Plateau in 1950, and numerous theses of The Ohio State University, in addition to Hunt's dissertation, describe the geology of the eastern part of the Gunnison Plateau (Hunt, Taylor, Babisak, unpublished theses).

In the Wasatch Plateau and near-by areas in central Utah Spieker and others have recognized all formations found in the western part of the Gunnison Plateau (Spieker and Reeside, 1925, p. 445, 448–449; Spieker, 1946, p. 123–124, 127, 130–133, 139). Spieker recently summarized his views on the geologic history of central Utah in a guidebook of the Utah Geological Society (Spieker, 1949).

Acknowledgments

The authors are indebted to E. M. Spieker, under whose direction the work was done, for advice and criticism both in the field and in the preparation of preliminary reports. William N. Gilliland and Siegfried Muessig offered invaluable consultation in the field in 1947 and in 1948, respectively. Carl A. Lamey identified the igneous rocks, and N. M. Denson read the manuscript.

Stratigraphic Systems

Jurassic System

Arapien shale.—The Arapien shale was defined by Spieker (1946, p. 123–124) as a formation with two members: (1) the lower Twelvemile Canyon member, and (2) the upper Twist Gulch member. In the west-central part of the Gunnison Plateau, the Twelvemile Canyon member forms rugged foothills which extend from the northern boundary of the area southward to about 1 mile north of Criss Creek. At this point it disappears beneath the Green River formation (Pl. 2, fig. 1). The Twist Gulch member crops out in the valley of the north fork of Little Salt Creek and is exposed continuously from there to the northern end of the Gunnison Plateau. This unit, which is less resistant than the overlying Indianola conglomerate, forms a distinct break in slope below the prominent cliff of conglomerate.

All the lithologic types recognized elsewhere in the Twelvemile Canyon member are present in this area (Spieker, 1946, p. 124). The lowermost exposed stratigraphic unit is a thin-bedded dark-gray limestone, with some inter bedded gypsum, which weathers light gray. It is closely folded and highly fractured. Hardy and R. E. Hunt have measured a minimum thickness of about 800 feet for this unit in Chicken Creek Canyon, east of Levan, Utah. A succession of gray shale and red shale with gypsum overlies the limestone in the west-central part of the plateau. An alternation of gray shale, sandstone, and

some limestone overlies this unit. The latter succession is estimated to be at least 1500 feet thick. The sandstone in a few places forms massive beds which commonly contain clay nodules, ripple marks, and some cross-bedding. The limestone is generally sandy or argillaceous and rarely exhibits mud cracks. Within this unit many intrusive masses of monzonite porphyry are found. A section, typical of the gray shale unit, was measured in Little Salt Canyon. It is nearly complete for this area, 1270 feet, and extends from the gray and red shale with gypsum to an angular unconformity with the overlying Green River formation. The Twist Gulch member consists predominantly of reddish-gray sandstone, red siltstone, and some red shale. It is relatively uniform throughout central Utah and no distinct units are recognized within the member. A section of 1839 feet was measured and is believed to be nearly complete, although the contact with the underlying Twelvemile Canyon member is covered at this point.

SECTION NO. 1, LITTLE SALT CREEK

Section of Twelvemile Canyon member of Arapien shale in Little Salt Creek

Green River formation

Twelvemile Canyon member of Arapien shale

	Thickness (Feet)
(33) Covered interval, some fine-grained sandstone	221.0
(32) Limestone, gray, dense, weathers yellow brown	5.0
(31) Sandstone, gray, calcareous, lower 18 feet forms cliff	62.0
(30) Sandstone, shale, and limestone; green gray to gray	23.5
(29) Sandstone, light green gray, calcareous, medium- to fine-grained, weathers yellow, gray clay nodules	4.5
(28) Limestone, light gray, arenaceous, thin-bedded	17.5
(27) Sandstone, light gray, calcareous, forms ledge	21.5
(26) Sandstone, shale, and limestone; poorly exposed, thin-bedded, more sandstone in upper part	113.0
(25) Sandstone, buff to light gray, weathers yellow brown, clay nodules, fossiliferous (*Ostrea* sp.), some ripple marks	6.5
(24) Shale, some sandstone, like Unit 13	74.0
(23) Sandstone, dark blue gray, medium- to coarse-grained, clay nodules; sandstone, buff to brown, in upper part	10.0
(22) Shale and sandstone, like Unit 13	7.5
(21) Shale, green gray, calcareous, fissile	3.0
(20) Shale, green gray, calcareous; sandstone, calcareous, thin-bedded	82.5
(19) Sandstone, green gray, calcareous, weathers yellow brown, some clay nodules, forms ledge	13.0
(18) Shale and sandstone, like Unit 13	8.5
(17) Sandstone, green gray, calcareous, 4- to 6-inch beds, forms massive cliff	17.5
(16) Sandstone and shale, thin-bedded, platy, like Unit 13	13.0
(15) Sandstone, gray, calcareous, 2- to 5-inch beds, clay nodules, forms prominent cliff	40.0
(14) Shale, blue gray, calcareous, clayey, weathers in rounded knobs	17.5
(13) Shale, blue gray, calcareous, fissile; sandstone, one-quarter inch beds	13.0
(12) Sandstone, gray, calcareous, weathers brown	1.0
(11) Limestone, gray, arenaceous; shale	14.5
(10) Shale, like Unit 9; sandstone; limestone, arenaceous	6.0
(9) Shale, blue gray, calcareous	46.5
(8) Shale, gray, poorly exposed	21.5
(7) Sandstone, gray, calcareous, fine-grained, thin-bedded, weathers in chips	26.0
(6) Sandstone, gray, calcareous, medium-grained, 1-foot beds, ripple marks and cross bedding	3.0
(5) Limestone, like Unit 2	39.0
(4) Limestone, gray green, argillaceous, mud cracks and possible trails	23.0
(3) Sandstone, gray, calcareous, cross-bedded	2.0
(2) Limestone, gray green, argillaceous, thin-bedded, platy, mud cracks	22.0
(1) Covered interval, shale, gray, calcareous; limestone, thin-bedded, platy; sandstone, thin-bedded	291.0
Total Twelvemile Canyon member measured	1269.5

The base of the Twelvemile Canyon member has not been found in central Utah south of the Wasatch Mountains. Near the mouth of Little Salt Creek, the Green River formation overlies the Twelvemile Canyon member with angular discordance; a red shaly bed at the

contact probably represents residual material accumulated on the old erosion surface. The Twist Gulch member seems to grade into the overlying Indianola group (undifferentiated) without observed erosional discontinuity. The Morrison (?) formation, which is recognized by Spieker and others along the southeastern margin of the Gunnison Plateau, has not been differentiated with certainty in this area (Spieker, 1949, p. 18).

Spieker (1946, p. 123) recognized that the Twelvemile Canyon member is probably equivalent to the San Rafael group. Hardy (1952, p. 26) believes that the thick gray shale unit of the Twelvemile Canyon member, which occurs elsewhere in central Utah, is a probable correlative of the Carmel formation, although definite paleontological evidence is lacking; the age of the Carmel is indeterminate between Middle and Upper Jurassic. The Twelvemile Canyon member of the Gunnison Plateau probably does not differ appreciably in age from that exposed elsewhere in central Utah. If the Twelvemile Canyon member is a correlative of the Carmel formation of the San Rafael group, the Twist Gulch member might be expected to be a correlative of the Entrada, Curtis, and Summerville formations of the San Rafael group (Hardy, 1952, p. 27–28). The age of the Twist Gulch member, therefore, is certainly Upper Jurassic.

SECTION NO. 2, LITTLE SALT CREEK

Section of Twist Gulch member of the Arapien shale measured in the north fork of Little Salt Creek

Indianola group

Twist Gulch member of Arapien shale

	Thickness (Feet)
(7) Sandstone and siltstone, like Unit 1...	154.5
(6) Conglomerate, red, pebbles and cobbles of quartzite (average diameter 1 inch); matrix red calcareous sandstone, forms massive ledge.....	10.0
(5) Sandstone and siltstone, like Unit 1...	104.5
(4) Sandstone, like Unit 2.....	15.0
(3) Sandstone and siltstone, like Unit 1...	1122.0
(2) Sandstone, red, coarse-grained to grit, forms massive ledge.....	10.0
(1) Sandstone and siltstone, red to red gray, calcareous, friable, thin-bedded; sandstone, coarse-grained in places.....	423.0
Total Twist Gulch member exposed.....	1839.0

Cretaceous System

Indianola group (undifferentiated).—This term has been designated by Spieker (1946, p. 127) to represent a thick series of conglomerate and sandstone which crop out north of Indianola, Utah, on Hjork and Dry creeks. The Indianola group (undifferentiated) crops out in the northeastern part of the area described in this report and is continuous into the northern part of the Gunnison Plateau, which is only about 15 miles from the type locality. Near Little Salt Creek the Indianola disappears beneath the North Horn formation and the Flagstaff limestone.

The Indianola group in this area consists almost entirely of massive cliff-forming conglomerate composed primarily of cobbles and boulders of dark blue-gray limestone and red and gray quartzite. Interbedded red sandstone and shale occur in the conglomerate and distinguish it from the overlying Price River formation. Beds of light-gray and buff sandstone, with layers of red shale, and gray or buff shale with interbedded light-gray calcareous sandstone are also present. A section of Indianola totaling 3600 feet was measured in Bear Canyon from the underlying Twist Gulch member of the Arapien shale to the angular unconformity with the Flagstaff limestone.

SECTION NO. 3, BEAR CANYON

Section of Indianola group (undifferentiated) measured at the southernmost outcrop along the west margin of the Gunnison Plateau

Flagstaff limestone

Angular unconformity

Indianola group (undifferentiated)

	Thickness (Feet)
(21) Conglomerate, red to gray, cobbles and boulders of red and buff quartzite and dark-gray dense limestone (maximum diameter 1 foot; average diameter 2–6 inches).....	579.0
(20) Sandstone and conglomerate, light gray to white; shale, red; conglomerate, pebbles and cobbles of red and yellow quartzite, dark-gray limestone, red-brown sandstone; sandstone, calcareous, coarse- to fine-grained; grit and sandstone, red and tawny, in upper part; shale, gray; unidentifiable plant fossils in upper 25 feet.....	401.5
(19) Shale, gray to buff; sandstone, light gray, calcareous.....	165.5

	Thickness (Feet)
(18) Shale, like Unit 16	4.0
(17) Sandstone, light gray to red, like Unit 10	19.5
(16) Shale, yellow gray and red	4.0
(15) Shale, red	31.5
(14) Sandstone, gray to brown, calcareous	4.0
(13) Conglomerate, red, boulders (average diameter 6 inches to 2 feet); shale and siltstone, red; sandstone, fine-grained	189.0
(12) Conglomerate, gray to red, cobbles and boulders of quartzite and limestone (average diameter 5 inches); sandstone, red, calcareous; shale, red	888.5
(11) Conglomerate, red, like Unit 9	173.0
(10) Sandstone, red, calcareous, fine-grained, forms ledge	8.0
(9) Conglomerate, red, cobbles (maximum diameter 6 inches) of dark blue-gray limestone and quartzite; shale, red; sandstone, red, calcareous, coarse- to fine-grained	148.5
(8) Conglomerate, like Unit 2	4.0
(7) Conglomerate, like Unit 3	16.5
(6) Conglomerate, like Unit 2	152.5
(5) Sandstone, red, calcareous; shale	24.5
(4) Conglomerate, like Unit 2, massive cliff in upper 20 feet	115.5
(3) Conglomerate, red, pebbles of limestone and quartzite; sandstone and shale, red, calcareous, in upper part	29.0
(2) Conglomerate, gray to red, pebbles and cobbles of dark blue-gray limestone, and quartzite, 3-to 5-foot beds separated by shale units	613.5
(1) Covered interval	30.0
Total Indianola group (undifferentiated) exposed	3601.5
Twist Gulch member of Arapien shale	

About 5000 feet of Indianola strata is exposed in the northeastern part of this area. In Chicken Creek Canyon, about 3 miles to the north, the uppermost conglomerates lack the abundance of limestone fragments which seem to be characteristic of the Indianola in the west-central part of the Gunnison Plateau. Hunt (unpublished dissertation) has measured at least 7000 feet of Indianola in the northern part of the Gunnison Plateau.

The Twist Gulch member of the Arapien shale appears to be conformable with the Indianola conglomerate. The upper part of the Indianola is not exposed inasmuch as the North Horn formation and the Flagstaff limestone overlap the Indianola with angular discordance. A relatively high area of Indianola conglomerate on the erosion surface beneath the unconformity is evidently related to the marked thinning of the Flagstaff limestone and the North Horn formation in the northern part of the area.

Fragmental plant remains were found 3000 feet above the base of the Indianola conglomerate in a 25-foot bed of buff sandstone. Schoff (1951, p. 626) found a marine zone of Colorado age about 9000 feet above the base of the Indianola in the Cedar Hills. Spieker (1946, p. 130) tentatively regards the beds below this marine zone as Upper Cretaceous. The Indianola group as identified in this area is therefore considered tentatively as Upper Cretaceous.

Price River formation.—The Price River formation was defined by Spieker and Reeside (1925, p. 445) with reference to strata which crop out in Price Canyon, northwest of Castlegate, Utah. The Price River formation is found only in the southern half of the area where it forms a prominent escarpment which trends northward (Pl. 2, fig. 2). Excellent exposures occur in each of the five major canyons which dissect the southern plateau surface. The formation also appears in the low hills near the southwest margin of the area.

The Price River formation in this area consists largely of conglomerate which contains pebbles and cobbles of quartzite and limestone. Interbedded with the conglomerate are medium- to coarse-grained sandstone lenses which weather to dark brown. Beds of red sandstone and shale, on the other hand, are common in the Indianola conglomerate.

The pebbles and cobbles of the Price River conglomerate consist dominantly of white, pink, and red quartzite. Smaller amounts of limestone occur chiefly as pebbles, although rarely the limestone is found as cobbles 6 or 8 inches across. The limestone is commonly either dark blue gray and dense or light gray and fine-grained. In some beds limestone constitutes about 10 per cent of the pebbles and cobbles. Pebbles of light-gray medium-grained sandstone and buff arenaceous limestone also occur in lesser amounts.

The greatest thickness of the Price River formation is exposed in Mellor Canyon, near the south boundary of the area, where about 1800 feet of strata was measured. The conglomerate crops out between a fault at the base of a prominent escarpment and the Flagstaff limestone which overlies the Price River with angular discordance (Pl. 2, fig. 3). Lenticular beds of light-gray sandstone, in addition to some conglomeratic sandstone, are conspicuous in Mellor Canyon.

SECTION NO. 4, MELLOR CANYON

Section of Price River formation and Flagstaff limestone on the north side of Mellor Canyon

	Thickness (Feet)
Flagstaff limestone	
(17) Limestone, gray, dense, thin-bedded to 18-inch irregularly bedded layers, blocky fracture, fossiliferous	111.0
(16) Limestone, gray, arenaceous; blue gray, dense; weathers buff yellow	44.0
(15) Limestone, gray, arenaceous, 6- to 18-inch beds, prominent ledges with cavernous weathering	64.0
(14) Limestone, gray to blue gray, dense, massive, irregular beds, chippy fracture, pisolitic structure, calcite in fractures, top bed forms prominent ledge	196.0
(13) Limestone, light gray, arenaceous, irregularly bedded, scattered pebbles of dark-blue fine-grained limestone, weathers mottled yellow	48.0
(12) Limestone, mottled yellow, red, and gray, arenaceous, irregularly bedded	10.5
Total Flagstaff limestone exposed	473.5
Angular unconformity	
Price River formation	
(11) Sandstone, light gray, conglomeratic, pebbles chiefly white quartzite (average diameter 3 inches)	80.0
(10) Conglomerate, light gray, pebbles and cobbles chiefly white, pink, and red quartzite (average diameter 4–10 inches)	115.0
(9) Sandstone, light gray, medium- to coarse-grained, conglomeratic, pebbles, chiefly white and red quartzite; sandstone, light buff, medium-grained, thin-bedded to 10- to 12-inch layers; grit, conglomeratic, cobbles (maximum diameter 6 inches)	214.0
(8) Conglomerate, light gray, pebbles chiefly white, pink, and red quartzite, matrix fine- to medium-grained	67.5
(7) Conglomerate, light gray, pebbles of quartzite; sandstone, buff, fine- to medium-grained, thin-bedded	42.5
(6) Sandstone, light gray, medium- to coarse-grained; limestone, light buff, finely crystalline or arenaceous	14.0
(5) Conglomerate, light gray, pebbles of quartzite and sandstone, matrix medium-grained	14.0
(4) Sandstone, light gray, fine- to medium-grained, beds 2–5 feet in thickness; sandstone, fine-grained, beds 2–4 inches in thickness or thin laminae; break in slope	31.0
(3) Conglomerate, gray, pebbles and scattered cobbles (maximum diameter 5 inches) of quartzite and light-gray limestone; sandstone, light gray, medium-grained, firm, beds at intervals of 5 to 10 feet	262.0
(2) Conglomerate, dark gray, pebbles of gray limestone (constitute about 50 per cent of unit in lower part) and cobbles (maximum diameter 6 inches) of brown quartzite and red banded quartzite; sandstone, conglomeratic, lenticular beds 2–3 feet in thickness; steep cliff	310.0
(1) Conglomerate, dark gray, pebbles and cobbles of pink, white, and red banded quartzite (maximum diameter 10 inches), gray sandstone, dense dark blue-gray and light-gray fine-grained limestone	164.5
Total Price River formation exposed	1314.5

In Mellor Canyon, near the south boundary of the area, the Price River formation occurs beneath the Flagstaff limestone with distinct angular discordance. Northward a wedge of the North Horn formation appears conformably beneath the Flagstaff and in angular relation with the Price River formation. These relations are clearly evident on the north side of Hell's Kitchen near the edge of the escarpment. The base of the Price River formation is not exposed in the west-central part of the Gunnison Plateau, although in Chicken Creek Canyon in the northern part of the plateau it occurs in angular discordance with the upper part of the Indianola conglomerate.

The Price River formation (Upper Cretaceous) has been recognized by Spieker (1946, p. 131–132) in the Wasatch and Gunnison plateaus. In the Wasatch Plateau the absence

of fossils in the upper part of the Price River formation and in the lower few hundred feet of the overlying North Horn formation leaves the exact age of the upper part of the Price River in doubt.

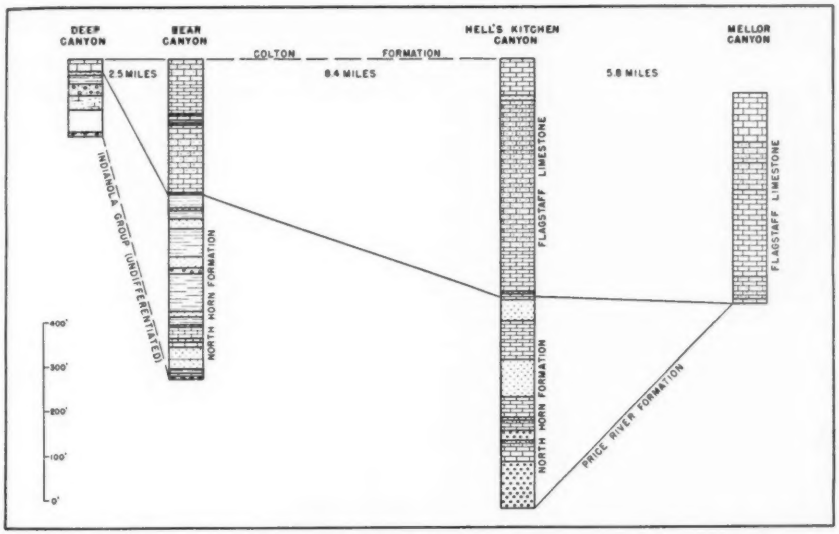

FIGURE 2.—CHART SHOWING LITHOLOGIC CHANGES IN THE FLAGSTAFF LIMESTONE AND THE NORTH HORN FORMATION

The massive conglomerate which appears in the southern part of the area is not continous along the outcrop with conglomerates which are considered Price River in other areas; however, it is believed to represent the Price River inasmuch as it occurs beneath strata which must certainly represent the North Horn formation. The angular unconformity between the Price River and the North Horn, in the west-central part of the Gunnison Plateau, is probably local and of no great time significance.

Cretaceous-Tertiary System

North Horn formation.—This formation was defined by Spieker (1946, p. 132) from exposures in the east-central part of the Wasatch Plateau. The unit was originally considered as the lower member of the Wasatch formation. The North Horn formation occurs in angular discordance with the underlying Price River formation and conformably below the Flagstaff limestone in the central part of the area. Excellent exposures are seen in Hell's Kitchen (Pl. 2, fig. 2) and Timber canyons. The North Horn is absent in the south where the Flagstaff limestone rests directly on the Price River formation. It is present, however, in the north where it is in angular discordance with the Indianola conglomerate. The best exposures of the North Horn are along the south side of Little Salt Creek Canyon.

The North Horn formation in general represents an alternation between fluviatile and lacustrine conditions (Spieker, 1946, p. 133). It includes an assortment of lithologic types which consist of variegated shales with sandstones, conglomerates, and fresh-water limestones (Fig. 2).

In the vicinity of Hell's Kitchen and Timber canyons, the basal unit of the North Horn is a massive conglomerate about 100 feet thick. Pebbles of white quartzite and dark blue-gray and yellow limestone are abundant. Overlying this unit is about 85 feet of light-gray arenaceous limestone with at least one bed of conglomerate lithologically similar to the basal clastic unit. In these canyons the upper part of the

North Horn consists largely of light-gray medium- to coarse-grained sandstones, and light-gray irregularly bedded limestones. Some beds of limestone in the upper part are mottled red, yellow, and lavender. This lithologic type is typical of the basal part of the Flagstaff limestone into which the underlying North Horn grades.

In the northern part of the area, the North Horn formation consists dominantly of red calcareous siltstone, some gray to red sandstone, conglomerate, and red shale. Mottled red and gray limestone identical with that found in the Flagstaff also occurs near the base. The North Horn formation locally pinches out over the Indianola conglomerate in this area.

Great variations in thickness characterize the North Horn formation. In the northeastern corner of the area, the formation is relatively thin and consists of yellow-brown friable sandstone, some shale, and conglomerate. It thickens rapidly southward to 422 feet in Little Salt Creek Canyon where it is dominantly red siltstone, gray sandstone, conglomerate, and red shale. In Timber Canyon there is at least 500 feet of strata which consist chiefly of gray limestone and sandstone with algal nodules, in addition to beds of conglomerate in the lower part. The North Horn thins southward from Timber Canyon and is not found in the south.

In the northern part of the area the North Horn formation overlies with angular discordance the Indianola conglomerate and the Twist Gulch member of the Arapien shale. The Price River formation, which normally underlies the North Horn in central Utah, is not recognized in this locality. The North Horn formation is in angular discordance with the underlying Price River formation in the southern part of the area. The contact between the North Horn formation and the overlying Flagstaff limestone is both conformable and gradational (Fig. 2). In the north the contact is arbitrarily placed above the dominantly red clastic sediments and below the strata which are dominantly limestone. In the south a considerable thickness of red arenaceous limestone is included in the basal part of the Flagstaff. The North Horn strata below this basal unit are characterized by light-colored massive beds of sandstone and limestone.

The North Horn formation lies below the Flagstaff limestone as recognized throughout the Gunnison Plateau and above conglomerates assignable to the Price River formation in areas where it is not in angular discordance with older formations.

In the eastern part of the Wasatch Plateau, the North Horn formation ranges from Upper Cretaceous to Paleocene in age. Spieker (1949, p. 27) believes that west of this area it is younger, but probably not younger than Paleocene.

Tertiary System

Flagstaff limestone.—The Flagstaff limestone was originally defined as the middle member of the Wasatch formation (Spieker and Reeside, 1925, p. 448–449) but was later raised to formational rank (Spieker, 1946, p. 135). The type locality is Flagstaff Peak in the southern part of the Wasatch Plateau. The Flagstaff limestone occurs over most of the north half of the area, although at most places it is covered by the Colton and Green River formations. In the southeast the Flagstaff forms prominent dip slopes on the plateau upland.

In the northern part of the area the Flagstaff limestone is characterized by mottled gray, yellow, and red argillaceous and sandy limestone (Fig. 2). The uppermost limestone bed, about 10 feet thick, is unusually persistent and traceable throughout the northern half of the area and beyond. It is a thin-bedded, light-gray, dense limestone which contains gastropods, ostracods, and veins of calcite. In Criss Canyon the upper half of the formation is dominantly yellow calcareous sandstone with some yellow limestone. The persistent limestone unit nevertheless occurs at the top of the section. A section, measured in Bear Canyon, is typical of the Flagstaff in the northern half of the area.

Section No. 5, Bear Canyon

Section of North Horn formation and Flagstaff limestone on the spur along the south side of Bear Canyon

Colton formation

Flagstaff limestone

	Thickness (Feet)
(46) Limestone, light gray, dense, thin-bedded, platy, fossiliferous (gastropods, ostracods), calcite in fractures	10.0+
(45) Limestone, light gray, arenaceous, scattered pebbles of limestone and quartzite, massive beds	7.0
(44) Covered interval, probably arenaceous limestone	10.0
(43) Sandstone, yellow, calcareous, forms cliff, vertical and lateral gradation to gray and brown limestone	10.0
(42) Limestone, gray, argillaceous, weathers in angular fragments, fossiliferous (gastropods)	19.5
(41) Limestone, like Unit 39	10.5
(40) Limestone, like Unit 26	2.0
(39) Limestone, light gray, arenaceous, dense, weathers yellow	8.5
(38) Limestone, red and yellow mottled, arenaceous, irregularly bedded	1.0
(37) Limestone, gray, dense, scattered pebbles of quartzite, weathers yellow	3.5
(36) Limestone, gray, arenaceous, pebbles near top, weathers yellow	10.5
(35) Limestone, gray and yellow, argillaceous, irregularly bedded	10.5
(34) Limestone, light gray and yellow mottled, arenaceous; limestone, gray, arenaceous, dense, in upper part	20.5
(33) Conglomerate, yellow, pebbles and cobbles chiefly red and white quartzite, matrix medium-grained	5.0
(32) Limestone, yellow and red mottled, arenaceous, argillaceous in part	8.5
(31) Limetone, yellow gray	3.5
(30) Shale, red and gray, clayey; limestone, red and gray mottled, in upper part	2.5
(29) Limestone, gray, arenaceous, beds 1½ to 2 feet in thickness	3.5
(28) Limestone, like Unit 26	3.0
(27) Limestone, gray, arenaceous, weathers mottled yellow, fossiliferous (gastropods), lenticular bed of conglomerate containing quartzite pebbles	15.0
(26) Limestone, gray and green gray, fossiliferous, forms massive cliff	15.0
(25) Limestone, red and mottled with gray yellow and gray green, arenaceous, irregularly bedded, nodular appearance	58.5
(24) Limestone, gray and mottled with yellow, fossiliferous (gastropods, algal nodules); limestone, arenaceous, near top; conglomerate, pebbles of limestone (average diameter ⅛ inch), single 4-foot bed	61.0
(23) Limestone, red and gray mottled, arenaceous	5.5
Total Flagstaff limestone	304.5
North Horn formation	
(22) Shale, red; siltstone, red, calcareous, in upper part	26.5
(21) Sandstone, light gray, calcareous, medium-grained, friable	8.5
(20) Shale, like Unit 7	16.0
(19) Sandstone, gray, calcareous, grades to siltstone	16.0
(18) Sandstone, light gray, calcareous	7.5
(17) Shale, red	3.0
(16) Siltstone, orange to red brown, calcareous, friable, irregularly bedded, fossiliferous (algal nodules)	2.0
(15) Shale, red and mottled with gray	10.0
(14) Siltstone, red, shaly in part	48.0
(13) Sandstone, light gray, medium-grained, calcareous, friable	14.5
(12) Siltstone, red, calcareous, irregularly bedded	5.5
(11) Sandstone, gray and red mottled, calcareous, thin-bedded	5.5
(10) Conglomerate, gray, cobbles and boulders of red, buff, and blue quartzite, and dark-gray limestone (diameter 2 to 12 inches), matrix medium-grained; sandstone, white, lenticular beds	16.0
(9) Siltstone, red, calcareous; shale, like Unit 7; sandstone, gray, fine-grained, calcareous, 1-foot beds	81.5
(8) Sandstone, gray to buff, like Unit 2; conglomerate, cobbles of blue and white quartzite and dark-blue limestone, lenticular beds	13.0
(7) Shale, dark red, clayey	20.0
(6) Sandstone, gray, like Unit 2	5.0
(5) Limestone, like Unit 3	27.5
(4) Sandstone, red to gray, calcareous	9.0
(3) Limestone, red and gray mottled, pebbles of gray limestone	10.0
(2) Sandstone, light gray, calcareous, weathers red; conglomerate, lenticular beds	27.0
(1) Covered interval, probably sandstone and siltstone	50.0±
Total North Horn formation	422.0
Indianola group (undifferentiated)	

In Timber Canyon the basal part of the Flagstaff is represented by about 190 feet of red irregularly bedded limestone. There are also several persistent sandstone layers, some of which contain conglomeratic lenses. Southward the basal part consists chiefly of irregularly bedded arenaceous limestone which commonly weathers to a mottled yellow, or red and yellow. The upper part of the Flagstaff between Timber Canyon and the southern margin of the area consists largely of light-gray limestone with some gray sandstone.

The greatest thickness of the Flagstaff limestone measured in the northern part of the area occurs in Criss Canyon where about 730 feet is found. Northward the formation thins to about 305 feet in the vicinity of Bear Canyon and thence to about 25 feet in the northeastern corner of the area. Southward from Criss Canyon it thins to about 650 feet in Timber Canyon and finally to about 500 feet in Mellor Canyon near the southern margin of the area (Pl. 2, fig. 3). Between Mellor and Timber canyons, the thickness of limestone above the basal red arenaceous limestone apparently is essentially uniform, although the basal unit is about 360 feet thick in Timber Canyon and only about 60 feet thick in Mellor Canyon. A section of the Flagstaff limestone from Mellor Canyon appears with the section of the Price River formation.

The red and yellow facies of the Flagstaff is developed notably in the northern part of the area and is not found immediately to the east or north in either the Gunnison or the Wasatch plateaus. The Flagstaff in the Wasatch Plateau typically consists of light-gray freshwater limestone and gray shale. It ranges in thickness from 300 to 1500 feet.

The Flagstaff limestone in this area is conformable with the overlying and underlying formations except in the northern part where it is locally in angular discordance with the underlying Indianola conglomerate (Fig. 2). Throughout most of the area, the contact with the overlying Colton formation is quite distinct. In the northwestern part, however, where the Colton formation cannot be separated from the Green River formation, the upper contact of the Flagstaff can nevertheless be readily placed. In the northeastern part of the area only a relatively thin section of Flagstaff overlies the Indianola conglomerate. Here the basal conglomerate of the Flagstaff consists dominantly of cobbles of red and green quartzite and only rarely contains limestone cobbles similar to those found in the underlying Indianola. Possibly this conglomerate represents the North Horn formation.

The Flagstaff limestone of the Wasatch Plateau was thought by Spieker (1946, p. 136; 1949, p. 32–33) to be probably Paleocene. Recent studies of the molluscan fauna by La Rocque (1951, p. 1457) indicate it is Paleocene and Eocene (?). The similarity of the stratigraphic succession of the Gunnison Plateau to that of the Wasatch Plateau suggests that the age is probably not greatly different in the Gunnison Plateau.

Colton formation.—The Colton formation was defined by Spieker (1946, p. 139) as equivalent to the original upper member of the Wasatch formation. At the type locality near Colton, Utah, it comprises 1500 feet of sandstone and red to variegated shale which intertongues westward with lacustrine beds of the Green River formation. The Colton formation is widely distributed in the eastern part of the area and also occurs near the southwestern margin. In the northwest, however, the Colton could not be differentiated from the Green River formation.

The Colton formation consists typically of gray, green, and red variegated shales; coarse-grained green-buff sandstone; light-gray medium-grained or brown sandstone; and some dark- to light-gray dense limestone.

About 600 feet of Colton occurs near the head of Criss Creek and about 350 feet near the southeastern corner of the area.

The Colton formation is conformable with the Flagstaff limestone below and the Green River formation above. The lower contact is quite distinct; the upper contact is probably not mapped everywhere at the same horizon as considerable intertonguing with the Green River formation is evident.

The Colton formation in the Wasatch Plateau is probably lower Eocene (Spieker, 1946, p. 139; 1949, p. 34), and it is unlikely that the age differs appreciably in the west-central Gunnison Plateau.

Green River formation.—The Green River formation was named by Hayden (1869, p.

89–92) for outcrops near Green River, Wyoming. Bradley (1931, p. 9) described the type section in greater detail. The Green River formation is present throughout southwestern Wyoming, northwestern Colorado, northeastern Utah, and central Utah. In central Utah it forms prominent hogbacks along the Wasatch monocline in Sevier and Sanpete valleys and also occurs throughout the southern part of the Gunnison Plateau.

In the west-central part of the Gunnison Plateau, the Green River formation occurs along the crest of the divide near the eastern margin of the area and also on the dip slope of the monocline between Little Salt Creek and Timber Canyon. In the northwestern part of the area it overlies the Twelvemile Canyon member of the Arapien shale.

The Green River formation in central Utah is largely fresh-water limestone and shale. In this area it is dominantly light-gray argillaceous thin-bedded limestone, gray shale, gray dense limestone, and light-gray coarse- to medium-grained sandstone. The tawny facies, which appears in the northwestern part of the area, contains more sandstone. It is stratigraphically above the typical Green River and is characterized by rapid vertical and lateral lithologic changes. The lower part of this facies is typically an alternation of gray-green calcareous shale and tawny medium-grained calcareous sandstone, with some gray dense arenaceous limestone and minor amounts of oil shale and conglomerate. Plant fossils occur in the shales and sandstones. The upper part is dominantly tawny dense limestone with some tawny medium-grained sandstone and interbedded gray shale, green-gray siltstone, and gray claystone. In the southern part of the area, the upper part of the Green River is represented by a conspicuous unit of green-gray shale, with a few interbedded layers of limestone, about 250 feet thick. At least 250 feet of light-gray limestone occurs above this unit. This sequence seems to be typical of the southern part of the Gunnison Plateau and of the Wasatch Plateau. The upper limestone may be the equivalent, in part, of the tawny facies of the Green River formation.

The total thickness of the Green River formation along the western margin of the plateau is thought to be more than 1400 feet. The tawny facies is nearly 1000 feet thick in Little Salt Creek.

SECTION NO. 6, LITTLE SALT CREEK

Section of the tawny facies of the Green River formation overlying the Twelvemile Canyon member of the Arapien shale in angular unconformity near the mouth of Little Salt Creek

Covered

Green River formation (tawny facies)

		Thickness (Feet)
(51)	Sandstone, tawny, coarse-grained; limestone, tawny	20.0
(50)	Limestone, tawny and mottled with gray; sandstone	25.0
(49)	Sandstone, tawny, calcareous; shale, gray green	25.0
(48)	Covered interval, probably gray-green shale, sandstone, and limestone	30.0
(47)	Sandstone, tawny, calcareous, friable, coarse-grained	10.0
(46)	Limestone, tawny, dense; shale and sandstone; fossiliferous (gastropods)	159.5
(45)	Sandstone, light gray to tawny, calcareous, friable	5.0
(44)	Sandstone, tawny, calcareous; conglomerate, pebbles and cobbles (diameter 1–5 inches) of dark-gray limestone and white and red quartzite	15.0
(43)	Limestone, green gray to yellow brown, arenaceous	10.0
(42)	Limestone, tawny; shale; sandstone	35.0
(41)	Sandstone, light gray, calcareous, friable	2.0
(40)	Sandstone, tawny, medium-grained	5.0
(39)	Limestone, tawny, irregularly bedded	40.0
(38)	Limestone, gray	10.0
(37)	Conglomerate, yellow brown, pebbles of limestone and quartzite; sandstone, yellow brown	5.0
(36)	Limestone, gray; claystone, calcareous	5.0
(35)	Shale, gray; limestone, tawny; sandstone; siltstone, green	30.0
(34)	Limestone, tawny, dense, calcite in fractures	2.0
(33)	Sandstone, yellow brown, calcareous; shale	10.0
(32)	Limestone, tawny; shale, gray	22.5
(31)	Sandstone, tawny, calcareous	7.5
(30)	Limestone, tawny, arenaceous; siltstone and shale, gray yellow green	60.0
(29)	Shale, like Unit 6; limestone, light gray, dense	25.0

	Thickness (Feet)
(28) Sandstone, gray yellow green, fine-grained, iron stain on weathered surface; shale	27.5
(27) Shale, green gray, clayey, calcareous	10.0
(26) Limestone, light gray	5.0
(25) Shale, like Unit 6	10.0
(24) Conglomerate, cobbles of red and gray quartzite (average diameter 3 inches), forms cliff	15.0
(23) Sandstone, light green gray, calcareous, fine-grained	1.0
(22) Shale and siltstone, light green gray and mottled with light brown, calcareous; limestone; sandstone; claystone, calcareous	35.0
(21) Limestone and sandstone, tawny; shale	50.0
(20) Sandstone, tawny, calcareous; limestone; shale, calcareous, plant fossils; conglomerate, lenticular bed 5 feet in thickness	84.0
(19) Shale, gray, calcareous; limestone, light gray, dense	28.0
(18) Sandstone, tawny, friable, coarse-grained to grit	6.0
(17) Sandstone and shale, plant fossils	13.0
(16) Shale, green gray; sandstone, gray; abundant plant fossils	11.0
(15) Limestone and sandstone, plant fossils (leaves, branches, twigs)	3.0
(14) Shale, like Unit 13 and Unit 19	6.0
(13) Sandstone, like Unit 15	3.0
(12) Shale, light olive gray, calcareous; shale, like Unit 13	4.0
(11) Shale, dark gray green, calcareous	1.0
(10) Sandstone, tawny, calcareous; shale, green gray	4.5
(9) Shale, gray blue green to gray; siltstone, arenaceous, calcareous	8.0
(8) Sandstone, gray and brown; limestone; shale	10.0
(7) Covered interval	25.0
(6) Shale, gray green and mottled yellow brown; siltstone; abundant plant fossils	3.0
(5) Limestone, gray; shale, green gray, clayey, calcareous	4.0
(4) Sandstone, tawny to gray, calcareous; shale, green gray, clayey, calcareous	3.5
(3) Sandstone, tawny, calcareous; sandstone, green gray, fine-grained, conglomeratic in part	4.0
(2) Conglomerate, gray, cobbles of green and red quartzite and limestone	1.0
(1) Covered interval	60.0
Total Green River formation (tawny facies) exposed	994.0

Twelvemile Canyon member of Arapien shale

The lower contact of the Green River formation with the Colton is conformable, although considerable intertonguing occurs. Where the Colton formation is not differentiated, the Green River appears to be conformable with the Flagstaff limestone although locally there must have been some erosion. Near the lower part of Salt Creek Canyon, the tawny facies overlies the Twelvemile Canyon member of the Arapien shale with angular discordance. South and east, the Crazy Hollow formation (Tertiary) overlies the Green River (Spieker, 1949, p. 36).

In the northern part of the area a few gastropods were found in that part of the Green River formation below the tawny facies. Fossil rushes, ferns, and a few deciduous leaves were found in the tawny facies along Little Salt Creek Canyon. Some ostracods were also found in the lower part of the tawny facies, and a small gastropod, *Helicina ?* sp. (A. La Rocque, personal communication), was found in the upper part.

Bradley (1931, p. 9) considers the Green River formation in Wyoming and northeastern Utah as middle Eocene. It is probably the same age in central Utah because the fossils found there do not differ from those described by Bradley (A. La Rocque, personal communication). Therefore, the Green River formation in this area is considered middle Eocene.

Igneous Rocks

In the northwestern part of the area small intrusive bodies of monzonite porphyry are found in the Arapien shale. The structural relationships of these masses to the surrounding sediments are generally obscured, although along Deep Creek and Little Salt Creek canyons two bodies are obviously sills. A third mass, which occurs in Maple Canyon in the north-central part of the area, appears to be discordant. The age of the intrusions is not known;

however, they have not been found to intrude beds younger than the Arapien shale.

The monzonite porphyry is composed of phenocrysts of hornblende in a groundmass consisting largely of orthoclase and plagioclase. The orthoclase is sericitized, and the plagioclase is extensively altered. Chlorite occurs as an alteration product of biotite and to a lesser extent of hornblende. Apatite is a principal accessory of the rock in addition to biotite, muscovite, and sphene. Specular hematite and magnetite are also abundant.

Structure

Regional Structural Relations

The Gunnison Plateau is in the area of structural transition between the Colorado Plateaus and the Great Basin (Spieker, 1949, p. 5; Fig. 1). The major structural feature of the plateau is a broad asymmetric syncline with the steeper limb on the east. This structure plunges southward from near the northern end of the plateau and extends beneath the alluvium of the Sevier Valley in the vicinity of Gunnison, Utah. Both the eastern and western margins of the plateau seem to be defined by major faults, although a conspicuous west-dipping monocline appears along part of the western margin. This monocline and related features are similar to the structure of the western margin of the Wasatch Plateau.

Major Structural Features

Monocline and associated structures.—A west-dipping monocline which involves the Flagstaff limestone and related formations extends along the western margin of the Gunnison Plateau from Timber Canyon northward to the vicinity of Buck Canyon, north of Little Salt Creek (Pl. 1). This structure is not found near the northern boundary of the area because the younger rocks involved in the folding have been removed by erosion, and the older folded sequence is exposed. In the vicinity of Buck Canyon, the Flagstaff limestone on the monocline changes in attitude from nearly horizontal to a dip of 31° W. South of Little Salt Creek, the Flagstaff and the North Horn dip 15° W. over the Indianola group and the Arapien shale (Pl. 2, fig. 1; Pl. 1, structure section A-B). The Flagstaff is nearly horizontal on the east side of Beckers Hollow, south of Criss Canyon (Pl. 3, fig. 2). Immediately to the west across Beckers Hollow, however, is a down-faulted block of the Green River formation with a westward dip. Between Criss Creek and Timber Canyon, the Flagstaff limestone and the Green River formation dip 13–17° W. although, north of Maple Creek, the Green River dips about 26° W. on the flank of the monocline. In a horizontal distance of 1500 feet down dip, at this point, the Green River changes abruptly to vertical and overturned attitudes. In the lower part of Mellor Canyon, near the western margin of the area, the Flagstaff limestone dips about 9° W. over the Price River formation (Pl. 1, structure section C-D).

The structure of the plateau upland is relatively simple. In general, the beds dip only about 2–6 E. and SE. Near the western margin of the area, a few normal faults of no great displacement seem to parallel the faults associated with the monocline.

The monocline is broken by a major fault in the vicinity of Timber Canyon and northward, near the western margin of the plateau (Pl. 1). This fault is well exposed on the north side of the lower part of Hell's Kitchen Canyon. At this point, a limestone characteristic of the upper part of the Green River formation directly opposes a reddish-gray conglomerate characteristic of the basal exposed part of the Price River formation. Stratigraphic displacement is about 1800–2000 feet. This fault parallels the outcrop of the Price River conglomerate south of Hell's Kitchen Canyon and apparently extends beneath the alluvium just north of Mellor Canyon.

Graben.—A major graben occurs on the shoulder of the west-dipping monocline in the western part of the area (Pl. 1). It extends from Criss Creek south to the boundary of the area. It is generally 1 mile, or slightly more, in width and about 11 miles long. At the northern extremity this graben trends nearly north. South near Maple Creek, the trend changes to about N. 24° E. and, near Hell's Kitchen Canyon, it again changes to N. 7° W.

In the vicinity of Criss Creek (Pl. 3, fig. 2) the fault along the eastern margin of the graben has a displacement of about 800–1000 feet. Green River beds west of the fault are found 760 feet below the top of the Flagstaff on the east. In the region of Timber Canyon (Pl. 3, fig. 1), the displacement may be as much as 1700–1800 feet where a limestone characteristic of the upper part of the Green River opposes the Price River conglomerate near the base of the escarpment. The displacement of this fault appears to decrease southward and at Mellor Canyon, near the southern boundary of the area, it is about 1000 feet as indicated by a down-faulted block of the Flagstaff limestone. Field relationships indicate

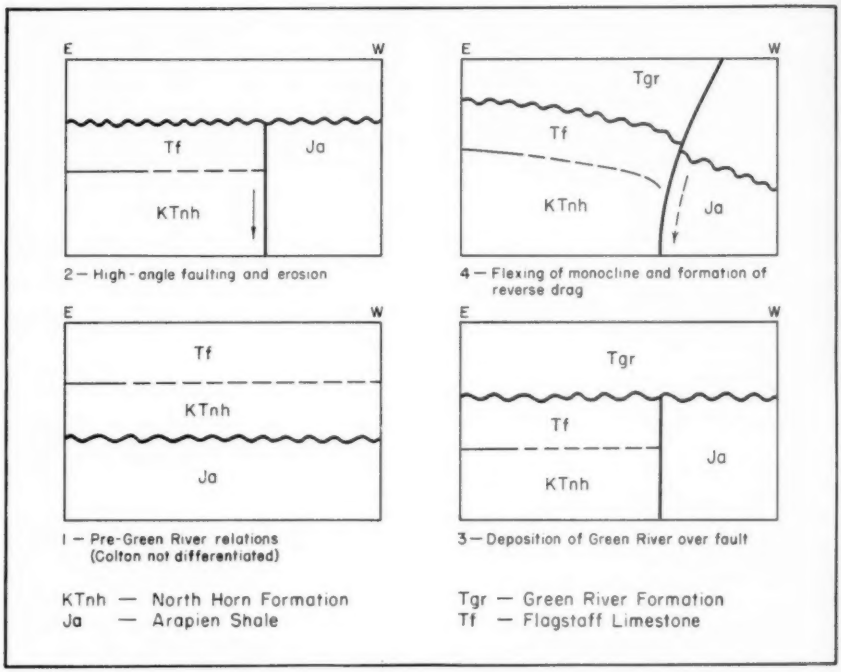

FIGURE 3.—SCHEMATIC DIAGRAMS SHOWING DEVELOPMENT OF PRE-GREEN RIVER FAULT

PLATE 2.—ANGULAR UNCONFORMITIES

FIGURE 1.—ANGULAR UNCONFORMITY, LOOKING SOUTHWEST, FROM LITTLE SALT CREEK

The tawny facies of the Green River formation, nearly horizontal, overlies the east-dipping Twelvemile Canyon member of the Arapien shale in the northwestern part of the area.

FIGURE 2.—ANGULAR UNCONFORMITY, LOOKING NORTH, ON THE NORTH SIDE OF HELL'S KITCHEN CANYON

Angular discordance is evident between the massive conglomerates of the Price River formation, in the lower part of the cliff, and the North Horn beds above, in the central part of the area. Both units dip toward the east.

FIGURE 3.—ANGULAR UNCONFORMITY, LOOKING NORTH, ON THE NORTH SIDE OF MELLOR CANYON

The Flagstaff limestone overlaps the east-dipping Price River conglomerate in the southern part of the area.

FIGURE 1

FIGURE 2

FIGURE 3

ANGULAR UNCONFORMITIES

FIGURE 1

FIGURE 2

GRABEN AND MONOCLINE

that the fault along the western margin of the graben is of less displacement.

The graben floor is 800–1000 feet lower topographically than the plateau upland and is fractured in a complex pattern. A remarkable series of west-dipping fault blocks of the Colton and Green River formations occur in the graben near Hell's Kitchen and Timber canyons (Pl. 3, fig. 1). In this area the displacements are about 150–200 feet, and the blocks dip about 34° W. Northward, similar small tilted fault blocks of the Green River occur on the graben floor.

Pre-North Horn structure.—In the northern part of the area, more than 7000 feet of pre-North Horn strata is present. These include the Indianola conglomerate and the Arapien shale. Except for the lower part of the Arapien shale, which is highly contorted, these units strike generally NE. and dip about 40° SE. In Little Salt Creek, however, the Indianola conglomerate and the Twist Gulch member. of the Arapien strike N. 60° E. and dip 40° SE.

In the central part of the area, on the north side of Hell's Kitchen Canyon, the Price River conglomerate dips eastward at a slightly greater angle than the overlying North Horn (Pl. 2, fig. 2). The Price River, at this point, dips about 5° E. In Mellor Canyon, near the southern boundary of the area, the Price River strikes nearly N.-S. and dips about 20° E. (Pl. 2, fig. 3).

Pre-Green River fault.—Near the mouth of the north fork of Little Salt Creek, the North Horn formation and the Flagstaff limestone overlie the Arapien shale in angular discordance on the east side of a normal fault; these units are not found west of the fault (Pl. 1, structure section A-B). The Flagstaff dips steeply toward the fault, but the actual truncation of these beds by the fault is obscured. It is unlikely that 455 feet of Flagstaff and North Horn could pinch out in a horizontal distance of 1500 feet. Probably these units were truncated and eroded on the west of the fault prior to Green River time (Fig. 3). The Green River formation, east of the fault, directly overlies the Flagstaff limestone since the Colton formation is not present in the area. West of the fault, the Green River formation is found in angular discordance with the Arapien shale (Pl. 2, fig. 1). The faulting is therefore dated as post-Flagstaff and pre-Green River. Reverse drag, apparent on the fault, might have resulted either from compression or subsequent movement on the fault.

Geologic History

The major geologic events, which are evident in the west-central part of the Gunnison Plateau, are similar to those outlined by Spieker for the geologic history of central Utah (Spieker, 1949), except for the pre-North Horn post-Price River angular unconformity, which is not found elsewhere.

Deposition of nearly 4500 feet of thin-bedded argillaceous marine limestone, gray shale with interbedded sandstone and gypsum, and red sandstone and siltstone occurred in Late Jurassic time. These strata constitute the Arapien shale. Coarse boulder conglomerate, 7000 feet or more thick, assigned to the Upper Cretaceous Indianola group, follows this succession without evidence of widespread disconformity. The lower portions of the conglomerate may be Lower Cretaceous, or perhaps Upper Jurassic. At least 7000 feet of conglomerate is found in the mapped area, making a total thickness of about 11,500 feet of strata below the angular discordance with the Price River formation.

The early Laramide orogeny (Spieker, 1947, p. 149, 152–155), widely recognized in central Utah, is dated by Spieker between middle and late Montana times. It is marked

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PLATE 3.—GRABEN AND MONOCLINE

FIGURE 1.—GRABEN, LOOKING NORTHWEST, FROM JUST SOUTH OF HELL'S KITCHEN CANYON

Note the fault blocks of the Colton and Green River formations on the graben floor between scarps of the Price River formation.

FIGURE 2.—MONOCLINE, LOOKING SOUTH, FROM CRISS CANYON

Small antithetic faults cut the west-dipping monocline on the right. The north-south marginal fault of the graben is at the left between a barren cliff and the brush-covered hills.

by an angular unconformity in the northern part of this area. The Price River formation, which appears to be the oldest unit above the unconformity, is not present here, and where it does appear in the southern part of the area the Indianola group does not outcrop. Nevertheless, the Arapien shale, the Twist Gulch member of the Arapien shale, and the Indianola group were extensively folded prior to the deposition of the Price River, and subsequent erosion produced considerable relief. The Price River formation was probably not deposited in the northern part of the area, although considerable thicknesses were deposited in adjacent areas to the northeast and south. A portion of the northern area remained positive through North Horn (Upper Cretaceous-Paleocene) time and through most of Flagstaff time (Paleocene-Eocene). Thus, in the northeastern part of the area, only the uppermost beds of the Flagstaff limestone were deposited over the Indianola.

In the southern part of the area, a thick section of the Price River conglomerate is overlain by the North Horn formation in angular discordance. The North Horn formation, however, wedges out toward the south so that only the upper Flagstaff beds are above the folded Price River conglomerate. This movement is later than the early Laramide orogeny established by Spieker (1946, p. 152–155) and appears to have preceded the pre-Flagstaff movement described by Spieker (1946, p. 155). Where the North Horn formation is present it is conformably overlain by the Flagstaff. Erosion, which followed the pre-Green River faulting in the vicinity of Little Salt Creek, locally removed the Flagstaff and North Horn. Thus the Green River formation overlies the older folded beds in places. Deposition of the Colton and Green River formations followed in a fresh-water lake.

The folding and faulting that affects the Green River formation cannot be dated from evidence in this area because beds younger than Green River are not present. Spieker believes that the flexing and faulting of the Wasatch monocline, east of the Gunnison Plateau, was initiated between late Eocene and Miocene (Spieker, 1949, p. 80). The Crazy Hollow formation of post-Green River age, however, is involved in the folding of the Wasatch monocline but is not present in this area. The similarity of the Wasatch monocline and the monocline on the western side of the Gunnison Plateau suggests the two monoclines are probably of the same age.

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PLEISTOCENE AND RECENT SEDIMENT FROM THE FLOOR OF THE NORTHEASTERN PACIFIC OCEAN

BY HENRY W. MENARD

ABSTRACT

Seven dredge hauls on seamounts, and nine cores of the sea floor show that glacial marine drift was deposited over several hundred thousand square miles of the northeastern Pacific during Pleistocene time. Some ice-rafted pebbles are covered by as much as 50 mm of manganese dioxide, suggesting a minimum age of 50,000 years.

Since glacial marine sedimentation ceased, blue mud has spread out from shore. The drift is covered by at least 127 cm of blue mud at a distance of 250 miles from shore. Southeast of Kodiak, Alaska, blue mud is covered by a thin layer of ash from the 1912 eruption of Katmai Volcano.

Ice-rafted pebbles surrounded by envelopes of well-sorted coarse sand are incased in pieces of a thick manganese crust from Gilbert Seamount. The pebbles settled on the relatively smooth surface of the manganese crust and were incased by the slow upward growth of the crust. Finer sediment, which settled from icebergs, was swept off the smooth crust except for well-sorted coarse sand which collected in the pockets in the crust surrounding the larger pebbles. The formation of the pockets and the envelopes of coarse sand is ascribed to the action of weak currents although the specimens were dredged from a depth of at least 3600 feet.

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INTRODUCTION

Expedition "Northern Holiday"

Expedition "Northern Holiday" was carried out in July, August, and September of 1951 by the Scripps Institution of Oceanography with the co-operation of the U. S. Navy Electronics Laboratory. The scientific program included physical and chemical oceanography and marine biology as well as marine geology, so only a part of expedition time was available for dredging and coring. The general results of the whole

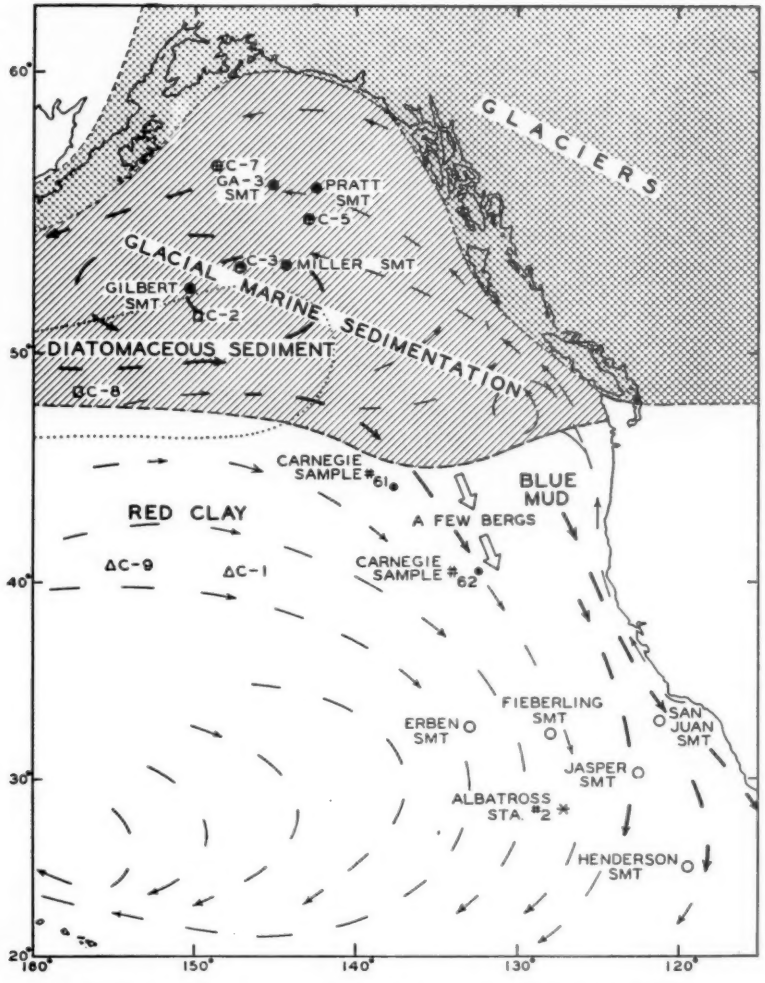

FIGURE 1.—DISTRIBUTION OF MARINE SEDIMENT IN PLEISTOCENE TIME

expedition and plans for future work on the samples and observations are summarized by Wooster (1951). During the expedition, four seamounts were dredged in the Gulf of Alaska, and the deep-sea floor was cored in nine places in the northeastern Pacific Ocean (Fig. 1). Many of the samples contain glacial marine sediments of the types widely distributed in the Atlantic Ocean (Peach, 1912; Bramlette and Bradley, 1940), the southern Indian Ocean (Murray and Philippi, 1908), and the Antarctic Ocean (Hough, 1950).

Previous Investigations

Pleistocene glaciers fringed the Gulf of Alaska, so ice-rafted sediment was expectable in the Gulf, although none had previously been reported. The U.S.S. TUSCARORA (Belknap, 1874) and the U.S.S. ALBATROSS (Murray, in

Agassiz *et al.*, 1902) took surface samples in the Gulf of Alaska, and similar samples have been collected by U. S. Coast and Geodetic Survey ships (Tyler, 1931). Neither these samples nor others collected in the more southerly parts of the northeastern Pacific contain sediment previously attributed to ice rafting, except for one about 600 miles west-southwest of San Diego, California (Agassiz *et al.*, 1902, p. 71). This extraordinary sample will be discussed in connection with the distribution of Pleistocene floating ice in the northeastern Pacific Ocean.

The "Northern Holiday" marine geology program in the Gulf of Alaska was designed to test the hypotheses of Menard and Dietz (1951, p. 1277–1285) as to the age and history of the seamounts. It was hoped that dredging and coring on seamounts would recover fossiliferous bedrock, but none of the bedrock dredged from side slopes was fossiliferous; however, the efforts of the party were amply compensated by the recovery of the dredge hauls of glacial marine drift, which are the main subject of the present paper.

Acknowledgments

The samples were collected on board the Scripps Institution of Oceanography research vessel HORIZON, James L. Faughn, Master. The chief scientists were John D. Isaacs up to Kodiak, Alaska, and Warren S. Wooster on the return to San Diego. All members of the scientific party and many of the crew co-operated in coring, dredging, and surveying; the writer is particularly indebted to Harris B. Stewart who bore a large share of the responsibility for dredging and surveying with tireless enthusiasm and good humor. The sea-floor cores were collected primarily for a study of Radiolaria by W. R. Reidel, who has made them available to the writer.

Identifications of heavy minerals and Foraminifera were made by E. L. Hamilton to whom the writer is indebted also for critical discussion of the manuscript in all stages of its development. Many of the writer's other colleagues of the Scripps Institution of Oceanography and the Navy Electronics Laboratory have contributed to this investigation. The comments and suggestions of M. N. Bramlette and R. S. Dietz have been particularly helpful.

METHODS

Equipment and Sampling Techniques

Dredge hauls on seamounts were made with a chain-bag dredge constructed at the Navy Electronics Laboratory for heavy-duty operations in the deep sea (Pl. 1, fig. 1). Steel plate half an inch thick and 10 inches wide frames the dredge mouth which measures 14 inches by 36 inches and resembles a box open at both ends. Behind the mouth is a bag made of 5-foot lengths of 1-inch chain wired together side by side and at the bottom. Another, longer bag of fish net with a half-inch mesh lies inside the chain bag and serves to retain finer material.

Pipe dredges were attached to the end of the chain bag by short lengths of chain. These simple dredges are short lenghs of 2½-inch pipe with attached cloth bags that retain sand.

The dredges were towed upslope at speeds of ½ to 2 knots with about 10 to 20 per cent more wire out than the depth indicated on the echo sounder. When the dredge was dragging over bedrock, the amount of pull indicated on a dynamometer was used to show if it was on bottom. Minor jerks and twitches of the wire could be felt even when dredging unconsolidated marine drift. The absence of such movement was taken as an indication that the dredge was above the bottom.

The corer used was similar to one developed by Fred B Phleger (Phleger and Parker, 1951, Pl. 2, fig. 1) but modified by the addition of heavier weights and a longer core barrel. This instrument obtains cores 1⅜ inches in diameter and up to 6 feet long. The cores were retained in the plastic tubes, which line the corer, and were not opened on board ship; however, a rough field description was made of the top and bottom of the cores before the tubes were corked. The depth of penetration, as indicated by sediment sticking to the outside of the coring tube, was determined in order to measure the shortening of the sample (Piggot, 1941; Emery and Dietz, 1941).

Laboratory Techniques

The plastic core liners and the samples were cut in half lengthwise. The tops and bottoms of the cores were sampled twice; one set of samples

was retained by W. Reidel, and the other set by E. L. Hamilton and the writer. The writer examined the cores with a binocular microscope and described any changes in color and grain size. Small samples were taken 5 or 10 cm apart or in every distinct layer. These samples were agitated and repeatedly decanted. The percentage of sand material in the original sample was estimated on the basis of the volume of the residue, and all grains of diameter of 1 mm or larger were counted and measured. Where this technique showed the presence of Foraminifera, a second sample was taken to be sieved because some tests were lost by the decantation. The samples from the tops, bottoms, and significant points in the cores were washed on 173-mesh bolting silk, and the percentages of Foraminifera and diatoms were roughly estimated by Hamilton and the writer. Hamilton subsequently identified these Foraminifera and also the minerals in the well-sorted sand layers in Core 2. Reidel examined the other set of samples for Radiolaria and diatoms. He has been kind enough to give the writer estimates of the percentage of diatoms based on a microscopic inspection of unwashed samples.

Summary of Results

The positions, depths, instruments, notes on collecting, and descriptions of the samples are summarized in Tables 1 and 2. Dredge hauls from Gilbert and Miller seamounts are shown in Plate 2. Schematic longitudinal cross sections of the cores appear in Figure 3.

Origin and Age of the Marine Gravel

Glacial-Marine Origin

The varied lithology, large particle size, poor sorting, and wide range of rounding of the sediment from the seamounts and from some cores leave little doubt that it was rafted into place. Floating ice is not now found in the Gulf of Alaska (Hydrographic Office, 1946, p. 3–18); but Pleistocene glaciers reached the continental shelf in the area (Flint, 1947, p. 223), and floating ice must have been widespread in the open sea. This ice appears to be the only agent capable of transporting so much coarse sediment so far. The sediment may be called *glacial drift* or simply *drift* according to both the original and present meanings of the terms (Flint, 1947, p. 102). Some fraction of the sediment may have been rafted by kelp (Emery and Tschudy, 1941). A floating clump of *Nereocystis* (Pl. 1, fig. 2) about 30 feet long was collected just east of Gilbert Seamount and roughly 300 miles from the nearest shallow water where it could have grown. The holdfast of one plant inclosed a fine-grained calcareous sandstone cobble measuring 80 x 50 x 40 mm. Several pieces of kelp were grappled from the waters of the open Gulf in other places, and many were sighted each day. If 1951 was a representative year, considerable material has been transported into the Gulf of Alaska by kelp since the last ice age.

Age of the Glacial Drift

No ice is found in the open waters of the Gulf of Alaska, so the glacial drift dredged from seamounts must have been deposited by ancient ice. The manganese film on the coarse drift confirms its antiquity because the rate of manganese deposition is very slow. Pettersson (1943) measured the rate of growth of manganese nodules by radioactivity analysis and found it to be about 1 mm/1000 years. If this rate is even of the right order of magnitude, manganese must not have been deposited in the Gulf of Alaska during most of postglacial time because the coating is much less than 1 mm thick. Evidence from other parts of the Pacific indicates that manganese deposition is intermittent or else that the long-term rate is two or three orders of magnitude slower than Pettersson's determination. For example, some of the seamounts of the Mid-Pacific Mountains have been under water since Cretaceous time (Hamilton, 1951), and Erben Seamount has been submerged since middle Tertiary time (Carsola and Dietz,

Plate 1. Chain-bag dredge and kelp-rafted sediment

Figure 1. Chain-Bag Dredge

Figure 2. Kelp-Rafted Sediment

Clump of *Nereocystis* with a cobble in one holdfast.

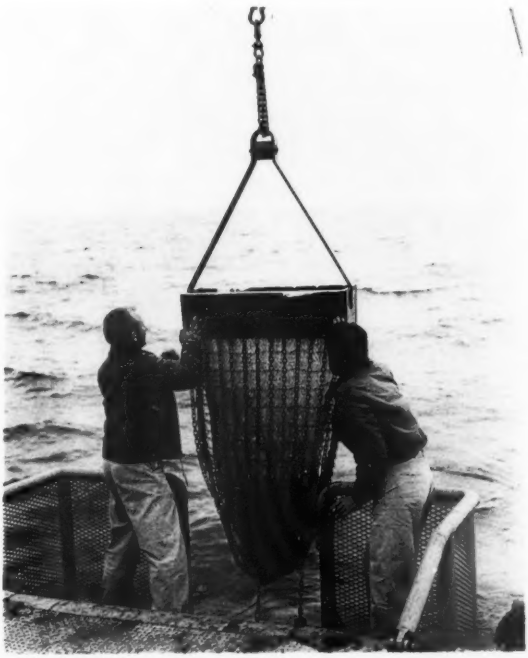

FIGURE 1

FIGURE 2

CHAIN-BAG DREDGE AND KELP-RAFTED SEDIMENT

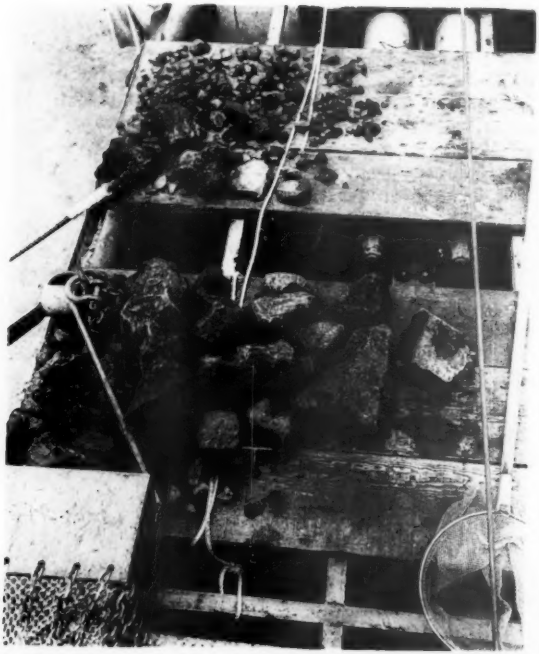

FIGURE 1

FIGURE 2

DREDGE HAULS

1952, p. 493), yet none appear to have a cover of more than 100 mm of manganese dioxide.

In addition to glacial drift, the dredge haul on Gilbert Seamount (NH D-1) included several fragments of manganese dioxide crust with a thickness of about 5 cm and an area as large as 15 cm square. The fragments are rather flat; the top is smooth except for shallow creases and low folds like those found on thick hide; the bottom grades into yellowish-white, weathered, earthy material rich in phillipsite and other alteration products. It appears that the fragments are parts of an extensive manganese crust formed on the weathered volcanic bedrock of the seamount. Scattered pebbles of sandstone, shale, basalt, and quartz, inclosed in envelopes of well-sorted coarse sand (Figs. 4, 5), are randomly distributed at all depths within the crust. The pebbles were rafted over the seamount from the time that the crust first began to form more than 50,000 years ago assuming the rate of deposition determined by Pettersson. However, thin laminations within the crust show that deposition was intermittent, so the total time involved may have been more than 50,000 years. Judging by the thickness of manganese crust on other seamounts which have a known minimum age, the crust on Gilbert Seamount may represent more than the whole of the Pleistocene. The observed relations between the erratic pebbles and the deposition of manganese on Gilbert Seamount permit only two conclusions:

(1) The unconsolidated drift is coated with a very thin layer of manganese dioxide and is relatively recent. (Presumably it was deposited either in the last glacial substage of the Wisconsin or during the whole of Wisconsin time.)

(2) The erratic pebbles incased in the manganese crust are much older than the unconsolidated drift. (They may have been deposited by floating ice during Wisconsin substages, or during the whole Pleistocene Epoch; or, conceivably, some may have been deposited by floating kelp or other means at any time.)

Pleistocene and Recent Sediment Distributions

Extent of Pleistocene Ice

Pebbles and coarse sand in Cores 2, 3, 4, and 8 indicate glacial marine sedimentation. In contrast, the red clay in Cores 1 and 9 contains no sand except authigenic manganese dioxide flakes. These cores, respectively 132 cm and 180 cm long, represent roughly 132,000 years and 180,000 years of deposition according to the rates of deposition given by Kuenen (1950, p. 384) so they would have penetrated at least to Wisconsin marine till if it were present. Apparently the southern limit of effective ice transport lay between the locations of the two types of cores (Fig. 1). Nearer the shore of the United States, the southern limit of Pleistocene floating ice is very poorly defined because the normal spread of terrigenous material from the continent obscures ice-rafted sediment. However, Revelle (1944, p. 32–33) finds that Carnegie Sample No. 61 (about 650 miles from shore, Fig. 1) is a gray clayey mud and believes "The color of this sample indicates terrigenous influence even though the distance from shore is great and the nitrogen content is not larger than that of other north Pacific clays." The terrigenous influence may be doubtfully referred to glacial marine transportation. To the southeast, Carnegie Sample No. 62 is a normal red clay (Revelle, 1944, p. 32) although it is roughly 250 miles closer to shore than No. 61 (Fig. 1).

One of the most unusual samples ever collected from the northeastern Pacific was dredged by the U. S. Fish Commission ship Albatross at Station No. 2 of Alexander Agassiz' third expedition in 1899–1900 (Fig. 1). According to Agassiz (1902, p. 71), the dredge contained red clay and manganese nodules as well as a few rocks thinly coated with manganese. The rocks were pebbles of:

hornblende andesite, rounded
pale-green serpentine, subangular

Plate 2.—DREDGE HAULS

Figure 1. Marine Drift from Gilbert Seamount
Figure 2. Volcanic Bedrock and Marine Drift
From the side slope of Miller Seamount.

TABLE 1.—DESCRIPTION OF DREDGE HAULS

Sample	Lat. and Long.	Topographic position	Depth in fathoms	Instrument and sampling notes	Description of sample
NH D-1	52–47 N. 150–05 W.	Gilbert Smt., near top	800–850	chain bag dredge and two pipe dredges, pipe dredges lost	About 500 pebbles, cobbles, and boulders, largest 20 x 30 x 33 cm. Most angular, some rounded. Various types of igneous, metamorphic, and sedimentary rocks. Several pieces of MnO_2 crust as large as 15 cm square and 5 cm thick. All large rocks and many small ones coated with a film of MnO_2. Sandstone, shale, basalt, and quartz pebbles imbedded in the MnO_2 crust.
NH D-2	53–32 N. 144–17 W.	Miller Smt., just below break in slope	550–600	chain bag dredge and two pipe dredges, pipe dredge bags torn and empty	One cetacean ear bone, one piece of MnO_2 crust 5 x 20 cm, 1 cm thick.
NH D-3	53–32 N. 144–17 W.	Miller Smt., on side slope	600–850	chain bag dredge and two pipe dredges, pipe dredge bags torn and empty, dynamometer pull reached 3½ tons	About 40 pounds of rocks, largely vesicular basalt with fresh fractures, unfractured surfaces coated with as much as 2 mm MnO_2. Some small pieces of MnO_2 crust 1 cm thick. Seven angular to rounded igneous and metamorphic rocks coated with a MnO_2 film, largest 10 x 20 x 20 cm.
NH D-4	56–20 N. 142–30 W.	Pratt Smt., on flat top	400	chain bag dredge and one pipe dredge	About 60 pounds of gravel and sand, angular to well-rounded, igneous, metamorphic, and sedimentary rocks, largest 10 x 12 x 20 cm. Most rocks coated with a film of MnO_2. Abundant forams in sand.
NH D-5	56–20 N. 142–30 W.	Pratt Smt., near break in slope	500	chain bag dredge and two pipe dredges	About 100 angular to well-rounded igneous, metamorphic, and sedimentary rocks coated with a MnO_2 film, largest 5 x 5 x 20 cm. One small MnO_2 chip 4 mm thick. Pipe dredges held 1 gallon of very poorly sorted sand and gravel with abundant forams.
NH D-6	56–10 N. 145–15 W.	Seamount GA-3*, near break in slope	500	chain bag dredge and two pipe dredges, chain bag empty	Poorly sorted sand and gravel, angular to rounded pebbles of igneous and sedimentary rocks, largest 5 cm diameter. Some forams in sand.
NH D-7	56–10 N. 145–15 W.	Seamount GA-3*, side slope	750–1000	chain bag dredge and two pipe dredges, dynamometer pull reached 3 tons	Freshly fractured volcanic breccia, unfractured surfaces coated with MnO_2 crust as much as 5 cm thick. A few rounded basalt pebbles imbedded in the MnO_2 crust.

* According to the numbering of Menard and Dietz (1951).

TABLE 2.—DESCRIPTION OF CORES

NH C-1 (40° 31.44′ N., 147° 58.5′ W., 2550 fathoms, 114 cm long, depth of penetration 132 cm)
Red clay throughout. No material of sand size except MnO_2 flakes.

NH C-2 (51° 34.5′ N., 149° 57.7′ W., 2450 fathoms, 145 cm long, depth of penetration 157 cm)

0–6 cm	Muddy diatom ooze, yellowish brown grading downward to dark yellow brown. Some sand, largest grains 1 mm. No forams.
6–37 cm	Light olive-gray clay. Some sand.
37–39 cm	Increasingly sandy.
39–40 cm	Well-sorted fine sand layer. Hamilton identifies the following minerals: quartz, orthoclase, albite (rare), calcite, glass, hypersthene, biotite, hornblende, augite, garnet, magnetite, apatite, zircon, and iron oxides.
40–56 cm	Diatom ooze, light olive gray grading to dusky yellow. A few coarse sand grains.
56–60 cm	Dusky yellow clay with fine sand.
60–70 cm	Dusky yellow-brown mud. Some sand, largest 5 mm. Diatoms abundant.
70–84 cm	Light olive-gray mud.
84–85 cm	Fine sand layer very rich in diatoms. Under a binocular microscope the sand is seen to be divided into three well-sorted sand layers each of different size. Graded bedding is not evident. Contains the same mineral suite as the sand layer at 39–40 cm except that albite is absent.
85–103 cm	Light olive-gray ooze with some sand grains.
103–105 cm	Transition to brown clay. Some sand grains.
105–112 cm	Dusky yellow-brown clay, darker at base. Some sand grains.
112–124 cm	Light olive-gray mud with some sand.
124–126 cm	Well-sorted fine sand. Contains the same mineral suite as the sand layer at 39–40 cm except that albite is absent, and oligoclase and clinozoisite are present.
216–145 cm	Light olive-gray ooze grading to more brownish downward. Some sand.
145 cm (bottom)	Well-sorted fine sand, very rich in diatoms. Largest grain 5 mm in diameter. No forams.

NH C-3 (53° 22.6′ N., 147° 13.6′ W., 2300 fathoms, 36 cm long, depth of penetration 63 cm)

Top	Diatomaceous sandy mud. About one third sand includes mineral grains and rock chips, largest 8 mm across. No forams.
10 cm	A 20-mm pebble in sand.
36 cm	(bottom) Sandy mud, largest grains 2 mm diameter. Diatoms rare, no forams.

NH C-4 (Top of Miller Seamount, 560 fathoms)
A few very small pebbles of granite, sandstone and quartz, and MnO_2 chips in core catcher.

NH C-5 (55° 07.8′ N., 142° 55′ W., 2030 fathoms, 127 cm long, depth of penetration 127 cm)

0–21 cm (top)	Reddish-brown mud, apparently the oxidized top zone of a blue mud.
Top	Very diatomaceous mud. No forams. No grains larger than 0.1 mm.
5 cm	A few forams.
10 cm	A few forams.
18 cm	Diatoms rare. Forams relatively common but only 10% of sample. Include *Globigerina pachyderma* (Ehrenberg), *G. bulloides* d'Orbigny, and *Globigerinoides rubra* (d'Orbigny). One 1-mm sand grain.
21–35 cm	A transition zone in color from reddish on top to blue gray below. Forams rare. No grains larger than 0.2 mm.
35–127 cm	Slightly bluish, medium dark-gray "blue" mud. Less than 5% nonclay. Grains as large as 0.2 mm rare. All grains larger than 0.2 mm mentioned below.
50 cm	One 1-mm rock fragment, one 0.5-mm rock fragment, but total amount of nonclay much less than 5%. A few forams.
60 cm	No diatoms. Very few forams, *Globigerina pachyderma* (Ehrenberg). One 1 mm grain.
80 cm	Very little silt, no sand. A few forams.
100 cm	A few very fine sand grains. Forams rare.
127 cm	(bottom) No diatoms. No sand. A few forams, *Globigerina pachyderma* (Ehrenberg), *G. bulloides* d'Orbigny.

TABLE 2.—*Continued*

NH C-6 (56° 00′ N., 144° 45.8′ W., 2100 fathoms, no core, apparent depth of penetration 120 cm)
The coring tube came up empty although a coating of blue mud on the outside of the corer indicated a penetration of more than a meter. Scrapings show no forams or sand grains.

NH C-7 (Axis of the Aleutian Trench, 56° 59.8′ N., 148° 49.4′ W., 2650 fathoms, 37 cm long, penetration not measured)

0–2.5 cm	Reddish-brown volcanic ash. Isotropic, index of refraction 1.487±0.001. Shards unaltered. No sand, 15% silt. Diatoms rare. No forams.
2.5–37 cm	Medium dark-gray "blue" mud.
37 cm (bottom)	No sand, a few per cent of silt. Diatoms rare. No forams.

NH C-8 (48° 37.5′ N., 157° 29.1′ W., 2750 fathoms, core 48 cm long, depth of penetration 48 cm)

0–48 cm	Reddish-brown sandy diatom mud from top to bottom. Much more diatomaceous at top than bottom.
Top	10% sand. Some MnO_2 grains. One basalt pebble 20 x 10 x 10 mm with a MnO_2 crust of perceptible but less than 1 mm thickness. One very fine-grained light-gray pebble 5 mm in diameter, coated with MnO_2. Very abundant diatoms. No forams.
10 cm	10% sand. Some MnO_2 grains.
13 cm	One 5-mm pebble coated with MnO_2.
20 cm	10% sand. MnO_2 grains more common.
30 cm	10% sand. One 2-mm quartz grain.
40 cm	5% sand. Largest grains 1 mm.
48 cm	15% sand. One pebble 8 mm in diameter, coated with MnO_2. Diatomaceous. Some Radiolaria. No forams.
(bottom)	

NH C-9 (48° 52.8′ N., 155° 13′ W., 2920 fathoms, 180 cm long, penetration not measured)
Reddish-brown clay throughout. No sand-sized material except MnO_2 flakes. Some diatoms and Radiolaria at top and bottom. No forams.

fine-grained sandstone of quartz, feldspar, epidote, and mica, angular
black chert with minute quartz veins, well rounded.

"How these pebbles, having all the characteristics of glacial-worn pebbles, came to be carried to this point is an interesting question. As the sharks' teeth and cetacean bones found in the red clay seem to indicate that since tertiary times there have been but very insignificant deposits at great distance from continental areas, similarly we may be tempted to assume that these pebbles, which could not have been transported to their present locality by any agency of the present epoch, were carried to it during the glacial epoch, the thin coating of manganese indicating how small have been the bottom deposits since the glacial period."

The possibility that the pebbles were kelp-rafted cannot be lightly dismissed because the area is so far south of other known glacial marine deposits. However, the uniform coating of manganese corresponds to the coating on the marine till dredged in the Gulf of Alaska, and the absence of uncoated pebbles suggests that the transporting agent is not acting at the present time. If the present general winter surface drift of the northeastern Pacific (Sverdrup *et al.*, 1942, Chart VII) is any indication of the drift during Pleistocene time, most of the drifting ice probably was trapped in the gyral in the Gulf of Alaska (Fig. 1). Nevertheless some of the larger bergs may have been swept far to the south in the cool waters of the California Current near shore.

Dredge hauls from five seamounts west and southwest of southern California (Fig. 1) show the rarity of glacial marine or other rafted sediment in the vicinity of Albatross Station No. 2. Many hundreds of pebbles were dredged from Erben and Fieberling seamounts (Carsola and Dietz, 1952, p. 484), but all were basalt. Jasper and Henderson seamounts were surveyed and dredged in June 1951 from the U.S.S. EPCE (R) 857, operated by the Navy Electronics Laboratory. All the scores of pebbles, cobbles, and boulders recovered were basalt. Likewise, dredging on San Juan Seamount (Shepard and Emery, 1941, p. 29) yielded no rock other than basalt. These dredge hauls should have contained some glacial marine sediment if

drifting ice was at any time common in the area. The absence of such sediment suggests that the berg or bergs that rafted the pebbles found at Albatross Station No. 2 were unusually large and long-lived, or else that the pebbles were not rafted by ice.

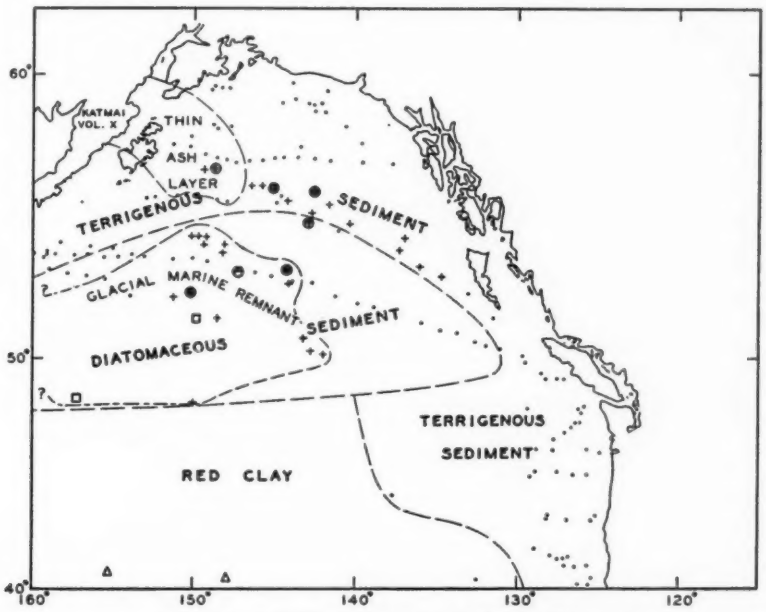

FIGURE 2.—PRESENT DISTRIBUTION OF MARINE SEDIMENT

Changes in Sediment Distribution

Changes in the areal distribution of sediment types on the deep-sea floor are produced by variations in the volume of sediment introduced by rivers, in the productivity of organic life, in current patterns (Arrhenius, 1950), and in particular by the presence or absence of floating ice. Some changes in distribution in the northeastern Pacific are shown by the Northern Holiday cores, and other changes can be inferred from cores, taken in the southern Indian Ocean, which lie in an analagous position relative to a glaciated area. The distribution of red clay, blue mud, and glacial marine sediment is relatively easy to infer compared to the distribution of diatom ooze and diatomaceous muds because the percentage of diatom frustules in a sediment varies with diatom productivity and solubility and also with quantity of sediment from other sources. For this reason no attempt is made to separate diatom ooze or diatom mud from diatomaceous red clay or blue mud. All sediment in which diatom frustules are an important constituent is grouped as "diatomaceous sediment."

The sediment distribution in late Pliocene time can only be inferred; presumably it included blue mud near shore, red clay in the deep open sea, and an east-west trending band of diatomaceous sediment near its present position (Fig. 2).

With the appearance of the first floating ice, marine drift began to spread over the northeastern Pacific and obscure the earlier sediment boundaries.[1] Moreover, according to Don Miller (Personal Communication) glacial marine sediment exposed along the northeastern shore

[1] The supply of floating ice must have varied from glacial to interglacial age with in the Pleistocene Epoch, but the whole epoch will be considered as one of dominant glacial marine sedimentation for ease of discussion.

of the Gulf of Alaska may be of Miocene age; if so, some Tertiary ice may have floated into the open sea. This drift is now found at or very near the surface of the sea floor only on the tops of seamounts and in the area of Cores 2, 3, and 8 (Fig. 2), but formerly it must have covered all the sea floor from the edge of the continental shelf to a southern limit at approximately 45° N. Latitude (Fig. 1). Very rarely bergs may have floated as far south as 28° N. Latitude.

The southern limit of the zone of diatomaceous sediment spread south with increasingly cold water, and the northern limit moved south because of dilution by rapidly deposited marine drift.

Melting back of the Pleistocene glaciers from the Alaskan continental shelf freed the waters of the northeastern Pacific from floating ice and cut off the supply of marine drift. As the water warmed, the diatomaceous zone moved north, so red clay began to lap over diatom ooze, and diatomaceous sediment was deposited on the southern zone of the drift. In the Indian Ocean, likewise, a few centimeters of diatom ooze have lapped over the outer edge of the marine drift which fringes Antarctica (Murray and Philippi, 1908; Schott, 1939).

While a thin layer of diatomaceous sediment encroached from the south, a much thicker flood of blue mud began to spread out from the continental shelves over the marine till. Core 5 did not penetrate marine till although it is 127 cm long, so a relatively thick layer of fine-grained sediment has spread for a distance of at least 260 miles from the edge of the continental shelf since glacial marine sedimentation ceased. For an order of magnitude calculation, it may be assumed that floating ice disappeared from the area 10,000 to 20,000 years ago, and that the minimum rate of sedimentation has been roughly 10 cm/1000 years even at such a great distance from shore. The rapid rate of sedimentation may be attributable in part to the large amount of rock flour supplied by melting glaciers. Tyler (1931) noted that the freshness of the minerals in his samples from the Gulf of Alaska indicated little chemical decay. The blue muds in Cores 5 and 7 consist of silt- and clay-size grains, but the material appears to be an unaltered rock flour with an admixture of pelagic organic detritus.

Ash from Katmai Volcano

The most recent change in deposition found in the Northern Holiday cores is shown by the occurrence of a layer of volcanic ash lying over blue mud in Core 7 (Fig. 3) taken near the axis of the Aleutian Trench east of Kodiak. This core slowly extruded from the coring tube when it was brought on deck. As it extruded, it was measured, and sections were placed in jars. A marked change in sediment type was noted 1 inch from the top, and the top inch was stored in a separate jar. At the time it appeared that the top layer was quite uniform as to grain size and color, but the inspection was superficial. Most of the sediment in the jars has since slumped into a shapeless mass so that detailed measurements of thickness or determinations of small changes in sediment type with depth in the core are impossible. Microscopic examination shows that the top layer is volcanic ash and that the sediment below is a blue mud like the mud in Core 5.

The general source of the ash layer is not difficult to find because a large number of active volcanos in the Alaskan Peninsula lie upwind relative to the constant west to east movement of the upper air. However, no other ash layers were observed in the underlying blue mud, so it would appear that perceptible thicknesses of ash were not ordinarily spread as far as 200 miles from the volcanoes to the position of Core 7. On the other hand no blue mud was observed on top of the ash layer so the eruption or eruptions which produced it probably occurred during the last few hundred years. (The rate of deposition of blue mud at Core 5 is roughly 1 cm/100 years.) An extraordinarily violent volcanic eruption within the last hundred years would appear to give the simplest explanation of the facts. The town of Kodiak lies roughly midway between the possible source volcanoes and the location of Core 7. Only two ash falls have been recorded there: ½ inch in 1867, and 10 inches in 1912. The latter fall is the only likely source of the ash in Core 7.

The well-known explosion of Katmai Volcano on June 6, 1912, blew about 5 cubic miles of rock into the air (Martin, 1913). Martin (unnumbered figure, p. 132) has drawn an isopachous map of the total thickness of ash from

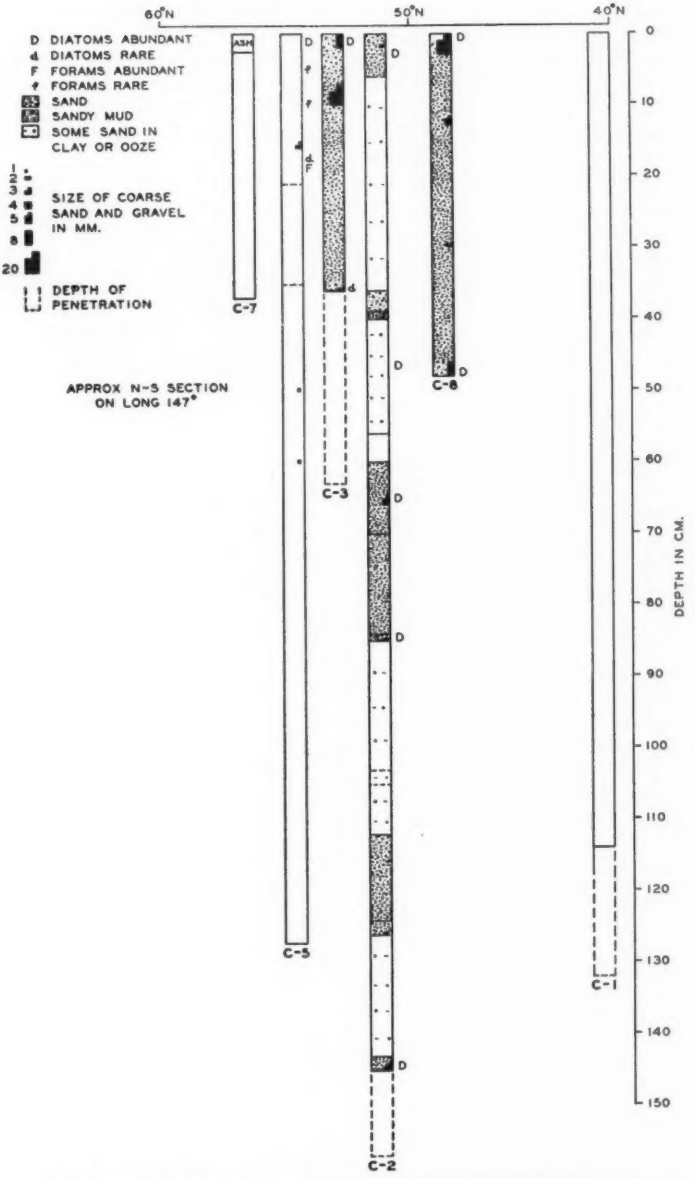

FIGURE 3.—DISTRIBUTION OF SAND AND GRAVEL IN DEEP-SEA CORES

the Katmai eruption. The contoured area is elongate toward the east-southeast because of wind transportation of the ash. The total thickness of the ash in this direction was 50 inches at a distance of 20 nautical miles from the volcano and 10 inches at 88 miles (Kodiak). Martin had no observation points at sea beyond Kodiak, but he placed the 1-inch isopleth about

175 nautical miles east-southeast of the volcano, and the ¼-inch isopleth about 195 miles east-southeast. The ash contained three distinct layers which varied in grain size with distance from the source. The top and bottom layers were gray or gray and white, and the middle layer was brown. This brown layer appears to correspond to the reddish-brown ash layer found at the top of Core 7 not only because of color and fine grain size but because both are composed almost wholly of volcanic glass with only a small percentage of mineral fragments.

The thickness of ash in Core 7 shows that the 1-inch isopleth was fully 213 nautical miles from Katmai. Bottom-sample notations on U. S. Coast and Geodetic Survey chart no. 8502 show brown mud for some distance east of the location of Core 7 (indicated by "Thin ash layer," Fig. 2) and gray mud east of the brown mud. The sampling devices used give a more or less composite sample. If it is assumed that ¼ inch of brown ash is the minimum thickness that would be detected in a composite sample collected in routine operation, the ¼-inch isopleth probably lies about 280 nautical miles east-southeast of Katmai Volcano. These observations show that the ash distribution was more widespread than Martin (1913) believed; however, his estimate was based on the rate of thinning of the entire ash deposit, consideration of the fine brown volcanic layer alone would suggest a different distribution. The material in Core 7 contains particles no larger than 0.1 mm in diameter, and Martin describes it as "impalpable dust" on land. Its settling velocity in air is about 1.5 knots. Martin (p. 176) did not find it as a separate layer closer than 61 nautical miles from Katmai Volcano in an east-southeast direction. At this point, Uganik Island, the layer was about 1½ inches thick so it probably existed somewhat closer to the volcano although it was unobservable because of the waters of Shelikof Strait. The ash was emitted from the top of Katmai Volcano at an elevation of about a mile, but it was seen to rise to great heights before spreading under the influence of the wind. Assuming a rapid rise to a total elevation of only 3 miles, the fine brown ash would require 2 hours to reach sea level. A wind velocity of only 30 knots would spread the ash to Uganik Island before any settled to the ground.

The fine brown layer at Kodiak is 4½ inches thick, although the town is 27 miles farther from Katmai than Uganik Island. It would appear that the wind produced a diffused cloud of fine ash and that only the edge of the cloud settled on Uganik Island whereas Kodiak was closer to the center. With these facts in mind, a distribution of brown volcanic ash on the sea floor beyond the limits estimated by Martin would be expected.

FIGURE 4.—POCKET IN A CRUST OF MANGANESE, AND THE PEBBLE FOUND IN THE POCKET

EVIDENCE OF DEEP CURRENTS

Scour on Gilbert Seamount

Water movement deep within the ocean is indicated by many lines of evidence:

(1) At all depths, the sediment on many topographic highs on the sea floor has been winnowed of fine sediment (Shepard, 1948, p. 303).

(2) Cretaceous rocks have been dredged from the tops of seamounts at depths as great as a mile—indicating prolonged nondeposition caused by the sweeping action of currents.

(3) Ripple-marked Globigerina ooze has been photographed on the top of a seamount at a depth of 750 fathoms (Menard, 1952).

The mode of occurrence of pebbles within the manganese crust from Gilbert Seamount provides a new line of evidence that sea water moves at great depth and that the movement is persistent; it also gives some clues regarding the current velocities required to sweep the tops of some seamounts free of sediment.

The pebbles are surrounded by an envelope

of well-sorted coarse sand containing some tests of Foraminifera (Fig. 4). Most pebbles and their sand envelopes are completely incased within the manganese crust. One pebble projects above the top of the manganese crust although it rests in a smooth-sided pocket in the crust. The pocket contains no sand, presumably because it fell out during the dredging. The pebble itself is imbedded in manganese at its base. Sand around completely inclosed pebbles was so loose that it could be poured out when the pockets were exposed by cutting in the laboratory. The manganese around the pebbles is thinly laminated so that the former surface of the crust at various times can be seen.

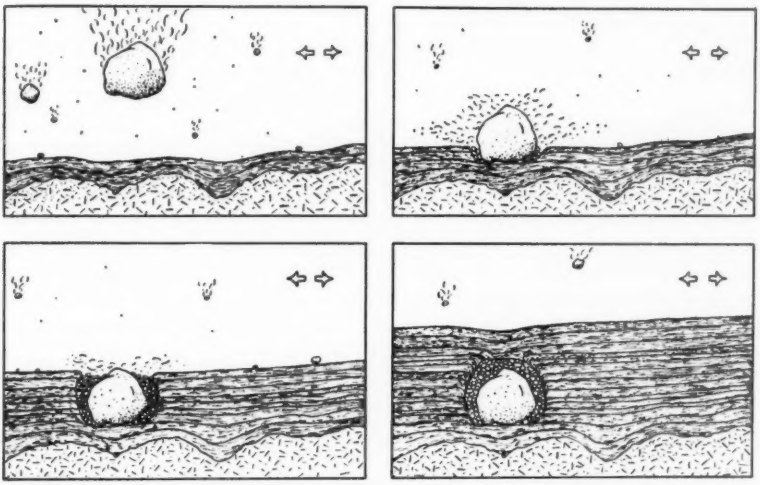

FIGURE 5.—SUPPOSED ORIGIN OF POCKETS IN A MANGANESE CRUST ON A SEAMOUNT

The supposed origin of the pockets in the manganese and of the sand envelope around the pebbles is shown in Figure 5. The pebbles of sandstone, shale, basalt, and quartz and also the sand and foraminiferal tests in the pockets suggest that a continuous (or frequent) rain of ice-rafted sediment fell on the seamount. The absence of sand layers or individual sand grains within the manganese crust shows that most of the rain of sand was swept off the smooth crust. Occasionally a pebble that was too large to be moved by the currents chanced to fall on the crust. Slow deposition of manganese produced upward growth of the crust inhibited for some reason—possibly because of increased turbulence connected with scour around the pebble. As the growth continued upward, a pocket formed around the pebble. except near the pebble where deposition was Sand which was swept across the crust collected in the pocket where the coarser grains remained although finer grains were winnowed out as though from a pothole. Eventually the crust grew so high that the top of the pebble no longer projected above the surface, and the pebble ceased to affect the flow of currents. The crust then grew laterally and sealed off the pocket.

Experiments show that pebbles with a nominal diameter of 14.5 mm (the size is not significant in the present discussion) and various shapes require mean current velocities between 14.5 and 42 cm/sec to put them in motion if they rest on the smooth painted bed of a flume (Krumbein, 1942). If 14.5 mm pebbles rest on other pebbles of the same size, however, the current velocity required to move them is about 150 cm/sec (Hjulstrom, 1935, p. 298). These observations indicate the important effect of bed friction on competent velocity for a given size of sediment.

The manganese crust is smooth although minor irregularities probably make the bed

friction higher than it is for painted wood. In the circumstances it would appear that the current velocity required to sweep coarse sand grains off the crust would be very small. The smallest pebbles found in any quantity incased in individual pockets in the manganese have a diameter of about 10 mm. The current velocity was not large enough to move these pebbles even on the smooth crust.

Sand Layers in Core 2

Core 2 contains four well-sorted sand layers separated by diatomaceous ooze and mud (Fig. 3). The upper three layers are only a few centimeters thick; the bottom of the core is in the fourth layer so its thickness is not known. The core was taken in 2450 fathoms of water about 320 nautical miles from the nearest land. The sea floor was relatively flat along the HORIZON track to north and south of the position of Core 2, but Seamounts GA-14 and GA-15 (Menard and Dietz, 1951, p. 1273) are only 40 to 50 miles to the northeast.

The layers contain a mineral suite (Table 2) which closely resembles the suites identified by Tyler (1931) in samples taken nearer shore in the Gulf of Alaska. The uniformity of the minerals in all parts of the Gulf argues for a common continental source rather than local sources on the sea floor. Ice rafting probably transported the sand to the vicinity of its present position. However, the sand layers in Core No. 2 were not emplaced directly from bergs because they do not have graded bedding which would be produced by the same differences in settling velocity which would separate the sand from the finer material settling from a berg. The sand layers must be the result of reworking of the ice-rafted sediment on the deep sea floor.

The nature of the process which caused the reworking is not apparent from an examination of Core 2. Similar sand layers in the deep sea have been tentatively ascribed to the action of hypothetical slumps and turbidity currents (Bramlette and Bradley, 1940, p. 16; Kuenen, 1950, p. 240; Phleger, 1951; Shepard, 1951). The sand layers in Core 2 provide no new line of evidence to support emplacement by turbidity currents. They give yet another proof that some process transports and sorts sediment on the deep sea floor, and in the absence of any other known process they are also tentatively attributed to turbidity currents.

CONCLUSION

Pleistocene changes in deep-sea sedimentation were widespread and profound just as they were on the continents. As till was spread over the continents by glaciers, so marine drift was ice-rafted over the northeasternmost Pacific Ocean to cover an area almost half as large as the United States. As periglacial action altered sedimentation on the fringes of continental glaciers, so changes in temperature and shifts in winds and currents altered pelagic sedimentation beyond the area of marine drift.

Post-Pleistocene changes in marine sedimentation have also resembled those on land. As the forests of pine and birch migrated north when the air became warmer, so the zone of high diatom productivity migrated north when the ocean water became warmer. As river gravels and dune sands covered glacial till, so blue mud and volcanic ash spread out from the continent and covered most of the marine drift in the northeasternmost Pacific Ocean.

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ORIGIN AND CLASSIFICATION OF SUBMARINE CANYONS

BY PH. H. KUENEN

ABSTRACT

Submarine canyons have generally been lumped together and a common explanation of their origin has been sought. The author suggests that some, named submarine ravines, are drowned river valleys (Corsica). The other extreme, the New England type, is the result of glacial turbidity currents. Daly's explanation of the California type as drowned subaerial valleys, smothered by sediment, and excavated by glacial and recent turbidity currents is accepted with some change of emphasis. This type is thus tentatively ascribed to the same turbidity-current mechanism, but erosion has exhumed some buried valleys, cleaned out parts of others, laid bare parts of old mountain slopes, and has even developed new channels. Shepard's new hypothesis of drowned river valleys kept permanently open by submarine processes, comes near to this picture and is accepted for the submarine ravines, but it cannot account for the features that indicate submarine erosion in the recent past off New England and California. His main argument against erosion by turbidity currents is the absence of scour on lake-delta fronts, but conditions in lakes must be highly adverse to erosion and hence this objection is eliminated. On the other hand Ewing and his associates give convincing evidence for the great importance of turbidity flow in the oceans.

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INTRODUCTION

Several attempts have been made to classify submarine canyons. It has been pointed out that those off the east coast of the United States differ somewhat from the California canyons, and the Mississippi Trough has generally been placed in a class apart. Small "gullies" have been contrasted with major

canyons. But none of these classifications has been generally accepted because there are all manner of transitions between them. The author tries to show that there are at least two extreme kinds of submarine canyons, apparently with entirely different modes of origin. The properties of one type should then no longer be used as arguments to explain the origin of the other. After the extreme types are established then the transitional types must be analysed to ascertain how and why they differ from the pure types.

The two extreme types are represented by the canyons off the east coast of the United States (the New England type) and those off the west coast of Corsica. The New England type is considered to result from turbidity flow during low levels of the Pleistocene (Daly's theory, 1936). Those off the west coast of Corsica are thought to be submerged land valleys only slightly modified by submarine processes. The same probably applies to most canyons around oceanic islands (the dominant process in Shepard's and Bourcart's explanations). The Tokyo and French Rivièra canyons are also drowned terrestrial valleys, but marine planation has cut back the coast several kilometers. The canyons further west in the Mediterranean are the New England type. It is suggested as a working hypothesis that off California an old land surface has been submerged, planed off, and silted up. During the Ice Age canyons were cut by turbidity currents, which cleaned out some old land valleys (Scripps, Carmel), opened up parts of other valleys (Partington, *e.g.*), encountered old, smothered mountain slopes along their walls (Monterey, Dume), and formed canyons with no relation to former land topography.

Acknowledgments

The author wishes to thank A. O. Woodford and F. P. Shepard for helpful criticism.

Submarine Ravines

Although the bathymetric survey of the Corsican ravines is far from adequate, some characteristics can be deduced. The deeply dissected mountainous land meets the coast in steep, cliffed headlands with deep intervening bays, an example of a drowned maturely dissected land surface (Fig. 1). The submarine ravines run up to the heads of these bays and form natural continuations of the land valleys.

Figure 1. Submarine Ravines on the West Coast of Corsica

They enter embayments and secondary indentations of the coast line. After Shepard (1948, fig. 79, p. 226).

Very little terrace is developed. The longitudinal slope of the ravine in the Golfe de Porto is about 1 in 9 (11 per cent) to 800 meters depth. The adjoining Porto River appears to have silted up its lower valley where the gradient over the last 4 kilometers is only 1 in 40 (2½ per cent). Farther inland it is 1 in 12 (8 per cent), and the valley walls are 1 in 2 or 3. The ravine in the Golfe de Girolata slopes about 1 in 6 (17 per cent) out to 800 meters depth. The adjoining river Tuara slopes 1 in 12 (8 per cent).

Allowing for some recent coastal and river

erosion and for deltaic sedimentation the morphology above and below sea level to some 800 meters depth is almost identical. The sparing data indicate that the drowned valley system does not continue below 1000 meters. Hence they end 1500 meters above the floor of the Mediterranean.

These submarine ravines are obviously drowned subaerial river valleys.

Although, because of lack of data, it cannot be ascertained how much submarine processes of erosion and sedimentation have modified these valleys, Shepard's (1952) recent hypothesis can be fully admitted for the ravine type of submarine canyon.

New England Type of Submarine Canyon

General Statement

The submarine canyons off the east coast of the United States are remarkably uniform in appearance (Fig. 2). Although they range from small valleys that hardly indent the edge of the continental terrace to the Hudson Canyon which heads 30 kilometers inside this edge, nevertheless the wall slopes, longitudinal profiles, slight sinuosity, and moderately dendritic pattern are much alike. Mainly for this type has an origin by the action of turbidity currents during the Pleistocene been claimed by Daly and the author.

The following morphological characteristics of the New England type of canyon must be accounted for by any hypothesis of their origin.

(1) The cross section is V-shaped. The wall slopes are around 1 in 4.5 (22 per cent) at the break in slope and gradually diminish in deep water, but may steepen again on steeper slopes farther out.

(2) The canyons follow straight down the dip of the submarine slope. By drawing lines tangent to the major convexities of the contours a generalized picture of this slope is obtained. Figure 3 demonstrates how closely the course of the major canyons conforms to the slope. Where the surface slopes straight down the canyons also run straight down to the deep-sea floor, but where the generalized contours show an outward bulge the canyons are deflected in an oblique course, thus conforming to the surface dip. Although the contouring (Fig. 4) by Veatch and Smith (1939, Plates) has been criticized by Shepard (1948, p. 208) and Kuenen (1950, p. 488) it appears safe to assume that the general picture, especially the course of the major valleys, is correct.

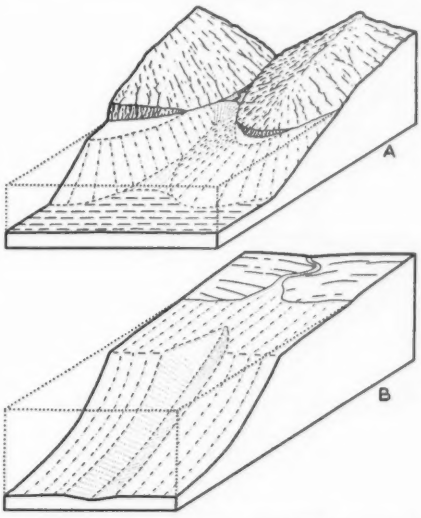

Figure 2. Block Diagram of the Two Principal Types of Submarine Canyons

A, the Corsican submarine ravines. B, the New England type.

(3) Inside the break in slope, where surface dip is no longer a controlling factor, the course tends to be deflected. The most striking examples are Wilmington, Baltimore, and Washington canyons.

(4) Bends are widely rounded.

(5) The floor grades outward continuously without counter gradients, even on very gentle slopes.

(6) The majority are steepest in the neighborhood of the head and gradually flatten off downstream.

(7) The gradient does not change at the break in slope of the continental terrace, except in the Hudson Canyon which shows a 2.8 per cent increase (Fig. 5).

(8) No "water falls" have been located.

(9) Tributaries enter at grade.

(10) No canyon is known to end half way down the slope. The Hudson canyon has been

traced to the Atlantic floor to a depth of at least 4500 meters, 500 kilometers out to sea. Its relief is greatest where it cuts into the steeper slopes such as the upper 2000 meters of the continental slope and the slope between 3100 and 3800 meters where it cuts a valley 550 meters deep and 5.5 kilometers wide in the ocean floor. In the almost flat area between 4000 and 4100 meters it exhibits evidence of meandering (Tolstoy, 1951, p. 443). Other canyons also reach the deep-sea floor. Possibly many fade out between 2000 and 3000 meters depth. Some smaller canyons and a number of tributaries start somewhere on the slope.

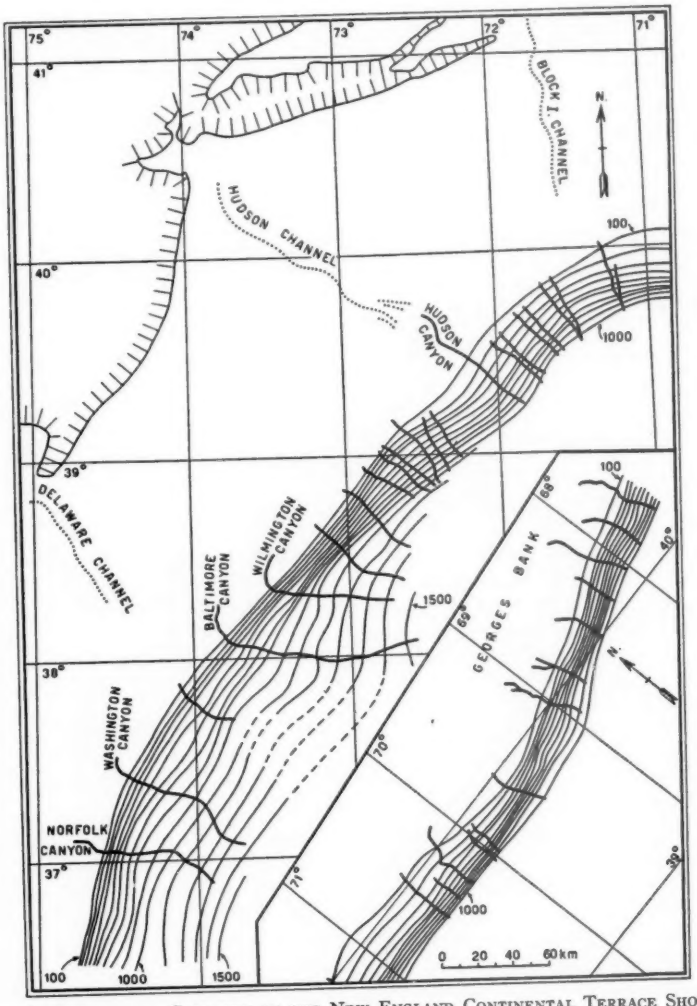

FIGURE 3. GENERALIZED CONTOURS OF THE NEW ENGLAND CONTINENTAL TERRACE SHOWING THE PRINCIPAL CANYONS

Depths in fathoms.

(11) Although the Hudson Canyon forms the continuation of the shallow Hudson Channel

on the shelf, there is no direct contact between the two (Kuenen, 1950, fig. 224). The Georges Bank canyons lack corresponding rivers; and for the southern canyons the link with rivers is doubtful to highly improbable.

the canyon at 2100 meters depth testify to the power of the currents flowing down the valley.

Seismic investigations of the continental terrace and a deep well drilled on Cape Hatteras show that this feature has been built up gradu-

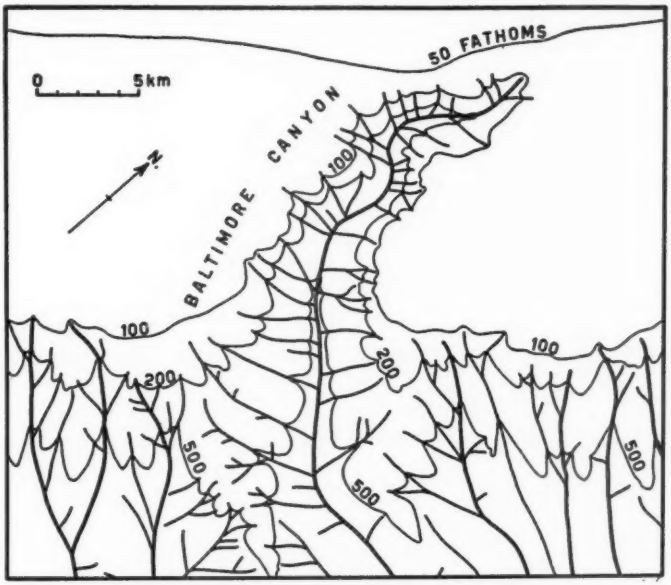

FIGURE 4. CHART OF THE BALTIMORE CANYON
After Veatch and Smith (1939, Pl. II B). Showing bends inside of the break in slope.

Dredging and coring (Stetson, 1936, Ericson *et al*, 1952) show that moderately or slightly consolidated Cretaceous to Pliocene rock crops out on the valley walls down to depths of 3500 meters. Therefore, although this does not exclude that the erosion may have started in the early Tertiary, deepening has continued at least until the end of the Pleistocene. As recent cover on the walls and sediment in the canyon beds usually show cold-water deposits below normal recent deep-sea sediment, erosion appears to have been particularly active during the Ice Age and to have ceased since. This is confirmed by the fact that around the lower end of the canyon in the "Hudson Delta" the graded deep-sea sands, containing shallow-water materials are thicker, coarser, and more frequent in the Pleistocene than later. This material was carried down along the canyon, because it is not found on the slopes beside the valley. Rounded hard-rock pebbles found in

ally since the Jurassic and now consists of unconsolidated Upper Tertiary some 300–600 meters thick, lying on semiconsolidated Lower Tertiary and Cretaceous, followed at several thousands of meters depth by the seaward-sloping surface of the basement. A few non-marine beds are intercalated in the Upper Cretaceous, but there is no evidence of emergence during the Tertiary. Emergence is also improbable, because consolidation is so little advanced.

Three possible modes of origin for the New England canyons are: (1) stream erosion, (2) ancient valleys modified by submarine processes, (3) submarine processes.

Stream Erosion

Recent cutting of the canyons by rivers can be definitely excluded. To expose the lower end of the canyons the sea level would have to be

lowered at least 2000 meters. The Hudson Canyon requires twice that amount. Physiographic, geologic, and climatic evidence exclude a great

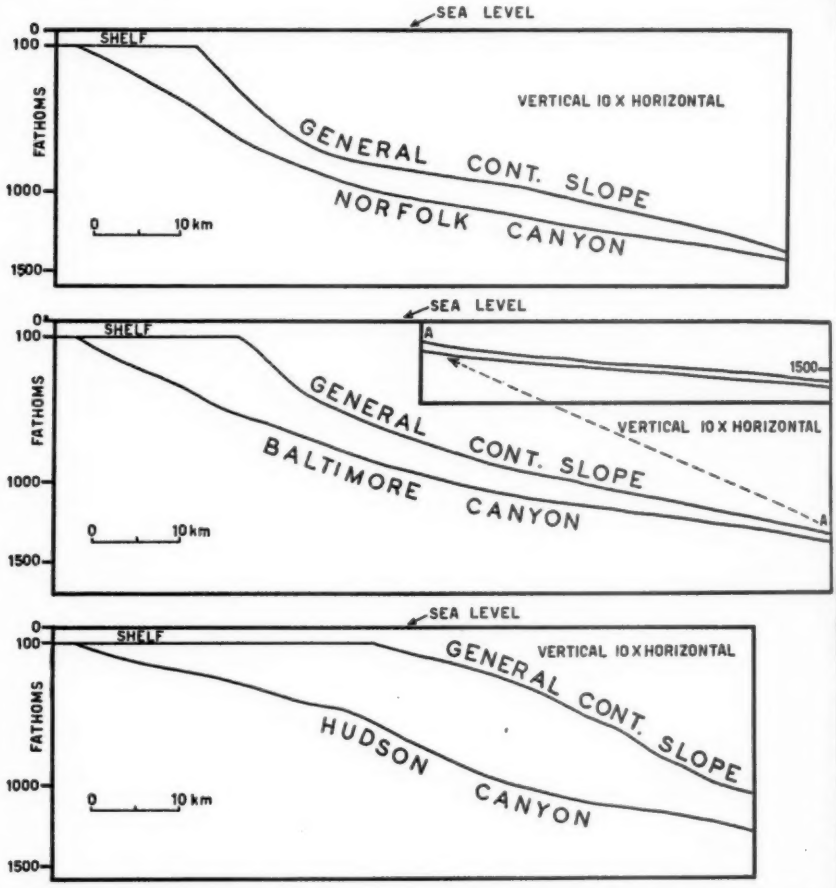

FIGURE 5. LONGITUDINAL PROFILES OF SOME NEW ENGLAND CANYONS WITH SHAPE OF CONTINENTAL TERRACE BASED ON GENERALIZED CONTOURS

Vertical 10× horizontal scale.

elevation of New England in recent times. A hingelike downwarping of the continental slope is ruled out because canyon profiles do not steepen at any hinge. No normal deltas are encountered at the lower ends.

Composite Origin

Shepard (1952, p. 84) has recently suggested a composite origin:

"Subaerial excavation of portions of the canyons at various times during the past when the margins were elevated above the ocean level; building of deltas on the outer slopes during excavation; submergence of the canyons, with some accompanying fill but with preservation of the canyons through slides and turbidity currents acting along the old canyon axes; reshaping of the old deltas, with some enlargement of shallow trenches characteristic of steep delta fronts; marine deposition on the old land surfaces into which the canyons were cut but concurrent maintenance of the canyons by slides, so that the canyon walls grew higher as the submergence continued; and, finally, reshaping of the canyon heads by Pleistocene sea level changes of a few hundred feet."

The advantages of this explanation over stream erosion are that subsidence may have been moderate and drowned valleys may have been extended seaward by turbidity currents either cutting unconsolidated sediment or re-

Shepard's hypothesis further implies that the canyon walls inside the edge of the shelf were subaerially cut except for the upper parts which were built up together with the shelf surface. Shepard's diagram (1952, Fig. 1, c) shows this

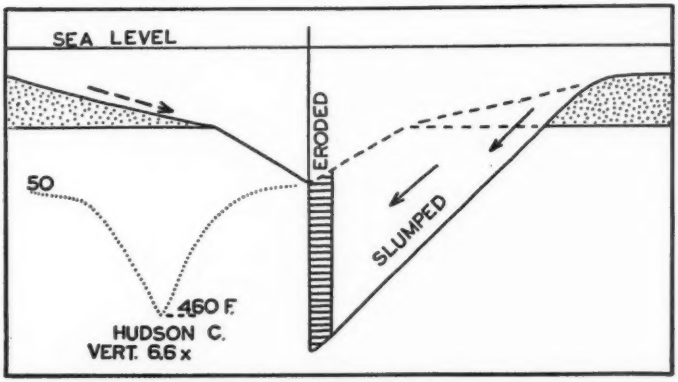

FIGURE 6. DIAGRAMMATIC SECTIONS OF SUBMARINE CANYON

Showing (on left) nick due to composite nature of canyon wall according to Shepard (1952, Fig. 1) and (on right) amount of turbidity-current erosion required to eliminate nick. One section of Hudson canyon (from Fig. 7) is added.

arding sedimentation along their course. The more remote the period of emergence the less forceful are rendered the arguments against a high stand of the continent. However, other difficulties are introduced.

The evidence against a great emergence of the terrace remain (no evidence from the Hatteras well, no consolidation). The close link between canyon direction and present dip of the continental slope indicates birth after the terrace reached its present shape. Even the edge of the shelf controls, by its position, the courses of the large canyons and the heads of countless smaller ones.

Several features oppose significant upbuilding after canyon cutting. The shelf is at the same level and reaches the same distance seaward on both sides of the canyons cutting into it. This is true even around the heads of those canyons which bend obliquely to the edge of the terrace (Fig. 4). But it is inconceivable that the supply of sediment could have been equal on both sides, for, wherever the supply was directed obliquely to the axis of the canyon, the shelf on the opposite side would be cut off from this supply by the canyon and hence could not have been built up and out to the same amount.

dual nature by a difference in slope for these two contrasting parts. Logically he draws the upper part, built below sea level, with gentler slopes than the gorge cut in exposed rocks (Fig. 6). But the charts show a continuous, straight, or slightly convex slope from shelf surface to canyon floor (Fig. 7). Slides or turbidity currents could not have steepened the upper parts of these slopes after deposition, because that would imply considerable deepening and inward lengthening of the narrow central part in recent times. While the supposed rivers were cutting their valleys the walls developed a certain angle. Later with less underwater weight these walls would not cave in until undercut to a steeper angle, as shown on the right in Figure 6. But even if this point is contested the nick cannot be eliminated without consolidation of the recent sediment and subsequent slump following undercutting. But undercutting by submarine erosion is the process which Shepard denies. Should one not expect at least a nick in the longitudinal profile where the subaerially cut valley fits onto the inward extension of submarine origin?

The upper parts of the walls are much too steep to have been built up to their present

slope directly. A slope of 1 in 4.5 cannot be expected under the given circumstances. Moreover, the continental slope shows that the maximum angle of repose during sedimentation is only 1 in 30 (Fig. 7).

ment younger than Eocene were removed from the terrace to show conditions before submergence, a very insignificant valley would remain to be attributed to river erosion. The same conclusion follows if one reasons that the

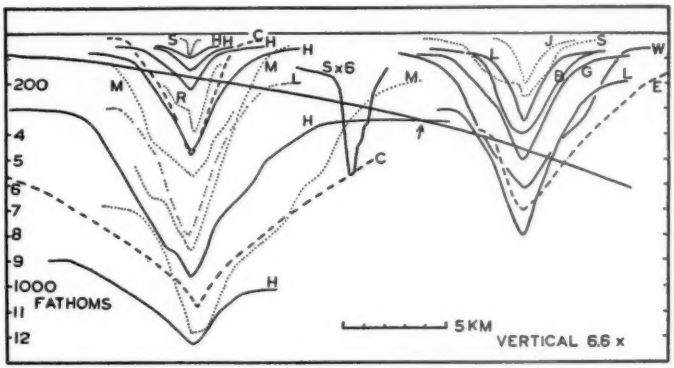

FIGURE 7. SECTIONS OF SUBMARINE CANYONS

New England: B = Baltimore, G = Gilbert, H = Hudson, L = Lydonia, W = Wilmington.
California: E = Eel, J = La Jolla, M = Monterey, R = Redondo, S = Scripps, C = Congo.
Note that among the California canyons only the shallow part of Scripps falls outside the range of the New England type. The undissected slope north of the Hudson Canyon is shown, the point of intersection with the canyon profile is marked with an arrow. Note the difference between slope of canyon walls and the terrace. Note absence of indication that river-cut and up-built parts join to make the wall slopes as claimed by Shepard.

If stream erosion excavated an appreciable amount of the canyons, the eroded materials should have built conspicuous deltas, now submerged on the continental slope. These are not found on the charts. If they are so old and insignificant as to be smothered by later sediments, the activity of the postulated rivers in forming the present canyons must have been mere scratching. There is no advantage in postulating this scratching for which there is no evidence.

Shepard (1952) has claimed that the lower reaches of submarine canyons are the result of non-deposition without any erosion. For the Hudson Canyon this would mean 550 meters of oceanic deposition since the turbidity currents started to keep the valley open. The great depth and absence of levees precludes the possibility that turbidity currents, instead of pelagic sedimentation, have built up the adjoining area.

On the estimated rate of deposition this oceanic sedimentation would require some 50 million years at the very least. But if all sedi-

terrace is built up swifter than the deep-sea floor. For very few of the canyons are cut significantly deeper than 600 meters into the terrace.

Obviously there is little advantage in postulating stream erosion if so small a proportion of the canyon cross sections can be attributed to this mechanism.

If, on the other hand, considerable erosion by turbidity currents is allowed for in the outer parts of the valley it is reasonable to postulate a larger amount on the steeper slopes. No river erosion is then required.

Submarine Processes

Several arguments oppose *sliding* as the sole cause of the canyons. Slides on land form U-shaped troughs with rounded heads, not narrow V-shaped valleys. The longitudinal slope of the canyons at greater depths is much too low. There is no accumulation of slumped material at the lower end equivalent in volume to the

missing valley fill. There is no concentration of canyons where the continental slope is steepest.

An array of features supports the explanation that the canyons were formed by *turbidity currents*.

Daly first called attention to this concept. He suggested that glacial lowering of sea level brought a vast amount of fine sediment on the shelf within reach of churning storm waves. The resulting heavy liquid would tend to run down the slope and in doing so could have excavated the canyons. The potential energy of sediment that stood high above the floor of the deep ocean was thus triggered off to cause the erosion. This original suggestion was later amplified by the author (Kuenen, 1947; 1950, p. 509). He showed, experimentally, the surprising power of turbidity flow that involved not only lutite but also coarse sand in suspension. Its competence to roll boulders along the bottom is remarkable. As the canyons have a much larger cross section and are much deeper than the currents can have been, only a narrow slot was actually abraded. The walls must have slumped in and evidently represent the maximum angle of permanent repose for the semi-consolidated terrace rocks. All these slumps should have formed turbidity currents, and most of the sediment required to set up the currents was provided by these slumps. In other words, given a lead the canyons eroded themselves.

The materials carried away would not form a true delta but a wide gently sloping feature gradually merging with the deep-sea floor. It was shown experimentally that this feature should be composed of graded beds between normal pelagic deep-sea deposits. Sharp bends, steep "water falls", and counter gradients should be absent. The canyons should follow, roughly, the dip of the present slope. A tendency to develop off the mouth of a river, or submerged river channel on the shelf, is explained by the vast amount of sediment brought out to sea by the rivers, that deepened their valleys during low sea levels.

Some of these features were already known to characterize the New England canyons (Stetson, 1936; Veatch and Smith, 1939); several others and some new arguments in favor of the hypothesis have recently been added by the staff of the Lamont Geological Observatory (Ericson *et al.*, 1952). It has now been established that the Hudson Canyon ends, as predicted, in a gently sloping cone of graded material with shallow-water components. It has been shown that the transporting agent carried pebbles to great distances along the valley but was of little importance on the adjoining slope. Slumps have occurred on the walls. Tertiary sediments including the Pliocene were cut. However, the activity became greatly reduced toward the end of the last lowering of sea level. The canyon is shallower and more sinuous on a gentle intermediate slope, but is deeper and straighter on the steeper slopes. This proves the erosive energy to be related to bottom slope, irrespective of depth. The importance of turbidity flow, in general, is convincingly proved by sampling in other areas of the Atlantic.

It also appears to the author that the volume of redeposited Pleistocene sediments will turn out to be so large that the major source will have to be sought in the erosion of the canyons, because the shelf and land could have supplied only a limited amount.

Finally the vast magnitude and almost incredible velocity of the Grand Banks turbidity current of 1929 gives eloquent testimony of potential energy incorporated in unconsolidated sediment high above the deep-sea floor (Heezen and Ewing, 1952; Kuenen, 1952).

The explanation of the delayed cable breaks following the 1929 earthquake has been questioned because of the high speed deduced. But, as Ewing has pointed out in discussions the velocity of dry avalanches, which are turbidity currents in the atmosphere, are no less remarkable. Dr. Th. Zingg of the Institut für Schnee- und Lawinenforschung kindly informed the author that he believes velocities of 200 km per hour are often reached and the limit may be in the neighborhood of 300 km per hour. The "nuée ardente" of Mt. Pelée in 1902, also a turbidity current in air, travelled at 540 km per hour.

These arguments in favor of canyon cutting by turbidity currents during glacially lowered sea level are formidable. The properties of the New England canyons are all consistent with this mode of origin. However, there is one controversial aspect. It is now firmly established that slumps can change to turbulent flow with

transporting power, but there is no proof that turbidity currents can erode.

The amount of material actually scraped away by the currents is small compared to the volume of the canyons. Most of it has slumped first, and it must be assumed these slumps were induced by undercutting and abrading of the floor. The seismic data, coring, and small maximum angle of rest (even under the stabilizing effect of reduced under-water weight) clearly demonstrate that the canyons were eroded in weak materials (Kuenen, 1950, p. 522).

The argument has been raised that no erosion takes place in Lake Mead and in the experiments. The low gradient in Lake Mead and the small scale in the experiments render these comparisons ineffective.

Another objection is that evidence of channel scour is insignificant in ancient graded bedding for which the action of turbidity currents is claimed. However, graded bedding was formed where the overloaded currents reached depositional velocities. It is more logical to claim that the vast volume of sediment in some beds indicates erosion higher up on steeper slopes.

The strongest argument is absence of scour on lake deltas. Shepard has argued that conditions for erosion on Swiss lake deltas are ideal because the material is unconsolidated and the slope is equivalent to or steeper than that in submarine canyons. Nevertheless the delta channels are nondepositional, the result of sedimentation on either side like the natural levees of rivers.

There are three reasons why this point is not convincing.

(1) The channel bottom is slightly but distinctly below the level of the delta surface beyond the levees.

(2) The small scale of lake phenomena reduces velocities and hence erosive power. A river diving below the surface has a very small discharge compared to the amount of heavy water that could flow off a shelf when a glacial storm had raised sediment.

(3) There are significant differences between the imagined processes on the shelf and delta sedimentation. On the shelf and slope erosion was restricted to short periods because turbidity flow occurred spasmodically, but then it was on a large scale, during occasional storms or earthquakes. In lakes, sedimentation and turbidity flow continue hand in hand. Normally the river water spreads along the surface and drops its load in an even blanket over a wide area. Even during the occasional times of underflow much sediment is wafted along the surface. Hence, any erosion by the turbidity currents is partly obliterated by "pelagic" sedimentation. In other words, blanket sedimentation was of no importance for the marine canyons, but is dominant on a delta front. By changing the position of its mouth a river starts new delta channels again and again and so cuts short the full development of turbidity gullies. Alpine rivers carry a much lower proportion of lutite and fine sand (presumably even less in the underflows) than was available for setting up turbidity currents on the continental slope. This must greatly reduce the efficiency of the turbidity flow.

The above review has shown that no satisfactory sequence of events can be postulated to show that the New England canyons were cut by rivers, either recently or in a remote past. But there is forceful evidence that they were cut by turbidity currents, almost exclusively during the Ice Age. The arguments against the ability of turbidity currents to cut into the terrace are not valid.

Intermediate Types of Canyons

General Remarks

If all submarine canyons were closely related either to the ravines or to the New England type the problem of their origin might be claimed as solved in principle. However, we are confronted with transitional types.

Canyons around oceanic islands are probably mostly the ravine type, that show steep slopes, irregular topography, entrance into deep bays, and, in many cases, mature dissection of the island (Shepard, 1948, Fig. 83). Subsidence of such islands is known to have occurred, guyots and atolls adding their testimony.

Mediterranean Canyons

The Mediterranean canyons off the southern coast of France are shown in Figure 8. In the east from Cannes to Marseille is a group of ravines, that enter into embayments of the

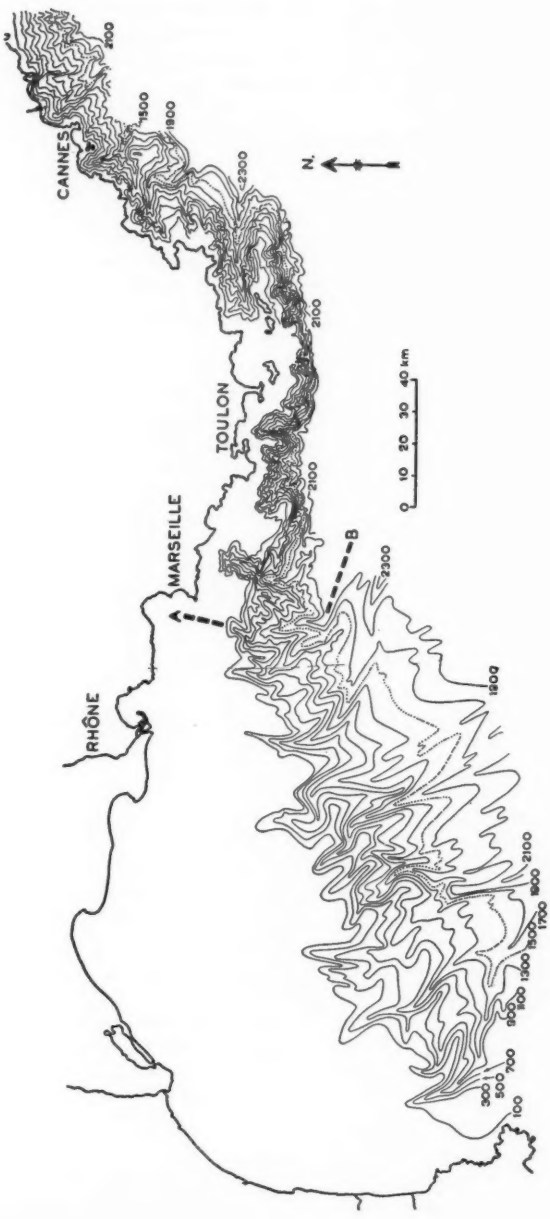

FIGURE 8. SUBMARINE RAVINES BETWEEN MARSEILLE AND CANNES
Contrasting with the submarine canyons west of Marseille (depths in meters). According to a chart by Bourcart (1950, Fig. 4).

coast line and show irregular topography, walls of ancient rock and abnormally steep longitudinal profiles (13 to 5½ per cent, compare Table 1). Typically, the direction is less rigorously controlled by the general dip of the

TABLE 1.—LONGITUDINAL PROFILE
PER CENT GRADE
(Shepard and Beard, 1938, p. 441)

	Head	Mid	End	Mean
East coast U.S.	7.3	5.5	4.8	5.5
West coast U.S.	10.0	4.0	3.4	4.8
Oceanic islands	21.3	14.2	11.3	13.8
Off large rivers	3.2	2.5	1.5	1.7

The general tendency of submarine canyons to show a steep head with a gradual decrease in grade toward the lower end is obvious. The great steepness of the ravines around oceanic islands is also evident. The California canyons are steeper at the head than the New England canyons. The lower gradient of the California canyons farther out is evidently due to the fact that they could be measured to their end, whereas measurements on the New England canyons could not be made more than half way down the continental slope.

continental slope than off New England. The shelf is extremely narrow in the east, but from Marseille to beyond Toulon it is well developed. Woodford kindly pointed out to the author that the stream gradients on land do not fit the longitudinal slopes of the Rivièra canyons. From these facts it may be tentatively concluded that marine and terrestrial erosion, and probably also some infilling of deeps, have modified the morphology significantly since subsidence ended (Rivièra type of submarine canyon). But waste has not been able to smother the topography effectively below 100 meters on the steep continental slope and must have been carried to the basin floor.

The Tokyo Canyon, which extends far into Tokyo Bay appears to belong in this group.

Bourcart (1950, p. 8) has claimed downwarping of the continental margin in the Rivièra to bring land valleys below sea level. This is possible, although we find no evidence of tilting (asymmetry of the large valley running out east to 2300 meters) or deformation of valley gradients. A purely vertical movement may also have taken place. A more reliable hypothesis must await detailed investigations of the shapes and composition of these canyons.

From Marseille westward the character of the submarine valleys alters abruptly. Longitudinal slopes drop to values normal for the New England type, and the topography loses its knobby character. Although the course as shown (Fig. 8) is more sinuous than off New England, the bends are sweeping, and the general direction conforms closely to the direction of the slope. These canyons are cut into the edge of a broad terrace that borders the flat depositional plain on land.

There is some indication on Bourcart's chart of levees along the lower reaches of these canyons, a common feature off southern California.

Bourcart has interpreted his dredging results as indicating two periods of erosion for the western group. However, a different interpretation is perhaps more likely. The canyons cut Lower Quaternary, Upper Pliocene, Lower Pliocene (?), and Miocene. The Pliocene is shown as filling an old, broad, flat-bottomed valley, but there is no proof that it does not extend laterally over a wide area as a normal unit below the Pleistocene. The canyon would then be entirely of Pleistocene age (Fig. 9).

The author suggests the same origin for these canyons as for the New England type.

California Canyons

Shepard's theory.—The most difficult problem is the California canyons which are better known than any other because of the amount of data assembled by Shepard, after years of field work. Hard rock, including granite, crops out on the walls at many places which leads Shepard to deny vigorously origin by turbidity currents. However, he gave up the stream-erosion theory and recently proposed a different scheme.

This explanation cannot be applied to New England canyons, but for California canyons it must be examined carefully and separately because it is superior to the pure turbidity-current hypothesis in explaining the hard-rock gorges.

The California canyons and their setting differ from the New England type. The shelf is much narrower, so that many canyons approach close to the shore, and a number head

directly off the mouth of a river. The canyons are more dendritic, the distribution more erratic, the control of direction by dip more flexible, and the section near the head in some instances much narrower with very steep to precipitous slopes. There is more evidence for rock from the walls including granite, basalt, conglomerate, and well-cemented sandstone. Longitudinal sections are less perfectly graded. The coastal ranges underwent great diastrophic revolutions in late Tertiary, possibly even in Pleistocene times. Highly raised marine terraces and deep valley fill testify to the instability of the continental margin. Table 1 shows that California canyons have gradients only slightly steeper than off New England. The cross sections, apart from a few narrow slots close to the shore, are remarkably similar (Fig. 7).

This setting is in sharp contrast with the New England canyons. But, apart from the nature of the wall rock, the canyons themselves differ more in degree than in kind.

Therefore although there are arguments in favor of submarine origin, there are fewer objections to the origin by drowning theory.

The lower endings of the canyons are now generally admitted to result from turbidity currents, although the details are uncertain.

The main arguments for invoking river action on land are the general similarity of the submarine canyons to land canyons and the hard-rock encountered in the walls. The first point bears little weight because turbidity currents can supposedly develop moderately dendritic valley systems with V-shaped cross sections and well-graded longitudinal profiles. However, the rocky walls present an insurmountable barrier to general acceptance of the turbidity-current hypothesis.

Shepard's new proposal combines subaerial and submarine processes, but still leaves unexplained the features indicating submarine origin, because he denies the ability of turbidity currents to erode even poorly consolidated deposits.

Daly (1936, p. 419) offered a different suggestion:

> "Here and there canyons and other types of valleys, cut by ordinary rivers in pre-Glacial time, have been drowned by strong, local subsidence of continental borders. Such old, rock hewn valleys, if they had not been quite filled with fine-grained detritus, would naturally have attracted the water loaded with sediment during the Glacial epochs. The resulting, localized currents would have been likely to remove some of the filling of each of the old trenches, and thus have revived the open-valley form more or less completely."

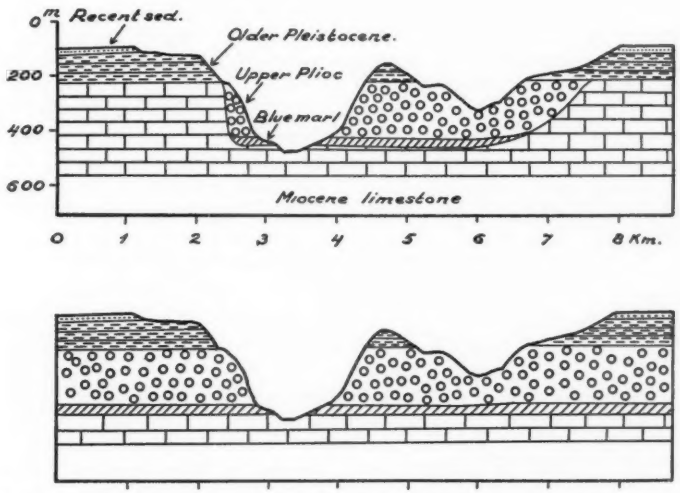

FIGURE 9. HYPOTHETICAL SECTIONS OF WEST MEDITERRANEAN SUBMARINE CANYON
Upper, according to Bourcart (1950, Fig. 2); Lower, an alternative reconstruction, which would require only one period of erosion.

The author would like to revive this suggestion because it offers a possibility of reconciling the apparently conflicting evidence.

Hypotheses compared.—(1) Both hypotheses start with the drowning of a dissected land surface along the present continental border. The highly unstable California coast has witnessed great vertical movements. Hence, drowning of deeply dissected topography with wide structural and narrow stream-cut valleys along the coast is a legitimate postulate. This may have occurred at different times in different areas.

But, while Shepard maintains that the canyons are land valleys kept permanently open by sliding and turbidity currents, Daly assumes burial and recent excavation. The smothering of such topography by sedimentation is known from the geological column. If a land of high relief has been submerged off California the smothering is indicated by the smoothness of the slopes off the islands and on submarine ridges. And, if a maturely dissected land topography has been obliterated on slopes away from continental supply of sediment, the infilling on the continental slope is probable.

Shepard contends that drowned valleys on gentle slopes were filled but remained open on steep ones. However, he presents no supporting evidence. There may be a general tendency in this direction, but we certainly find some steep slopes with no or only widely spaced canyons (from Ventura southward) and some gentle slopes with a group of narrowly spaced canyons (*e.g.*, Arguello group). The relatively large number of canyons heading off river mouths, where smothering sediment should be particularly plentiful, is difficult to explain by Shepard's postulates.

(2) Not a single canyon passes inside the coast line although a dozen head within a stone's throw of the beach. This is very strong evidence that erosion did not start until recently and was then limited to the marine environment. Ancient valleys cannot have been restricted in length by the much younger present coast line. Had they passed farther inland the slumping process invoked by Shepard should have been able to keep the portion indenting the shore open.

(3) Woodford (1951, p. 833) has shown that the almost straight longitudinal profile of the Monterey Canyon down to 2000 meters is in marked contrast to the logarithmic curve of mature inland rivers. Hence it is not a normal river valley. Moreover, the compound Salinas-Monterey profile shows a pronounced break on weak sediments at the shore line. This holds for other canyons heading at river mouths and shows the submarine canyons are recently developed features.

(4) The control of canyon direction by the present dip of the continental slope is obvious. Figure 10 shows that off the represented part of the coast the canyons are not at right angles to the general trend of the coast line but tend in a marked degree to cross the generalized contours of the slope at right angles. In Figure 11 the convergence downstream in embayments of the slope is in sharp contrast to the parallel course on the straight parts above. Figure 12 demonstrates the same features; the only marked exceptions are Partington Canyon, which runs parallel to the contours to join Sur Canyon, and the direction of Carmel Canyon. A third exception off California is Santa Cruz Canyon.

This control by the present slope militates against a remote origin, because the submarine relief is certainly no older than late or post Tertiary. For instance, it can be shown that at Mugu Point the Miocene sediment was supplied from the south precisely from the direction in which Mugu Canyon now runs down into the Santa Monica Basin.

(5) Bends in the submarine canyons are more widely rounded and the general system more nearly parallel than the land river valleys. The marked difference is brought out by comparing Figure 13 with the charts of the area off the California coast. Figure 13 shows the pattern of the valley system on part of the Pomona quadrangle. A neighboring area was used by Shepard (Shepard and Emery, 1941, Fig. 31, p. 85) to illustrate the close analogy between subaerial and submarine canyons. Actually it is obvious that even in this favorable area the general slope from the Gabriel Mountains to the coast exerts much less control on the direction of the valleys than does the continental slope. The spacing is much narrower than on the continental slope.

(6) Land valleys tend to be much more diversified in shape than are the California sub-

marine canyons. A point of importance is the absence in the submarine canyons of certain features, especially broad, horizontal bottoms. If old valley systems had been submerged one would expect that some showed such bottoms to begin with and retained them, and that others developed counter gradients or were much modified by later vertical and horizontal fault activity. Had later submarine processes, possessing no erosive power, been able to keep such valleys open continuously, basins should have developed with broad flat floors, sloping only slightly seaward, where sediment was ponded up-slope from diastrophically produced, hard-rock barriers. High "water falls" and sharply off-set valleys with abrupt ends might also be expected. Such features appear to be absent or of minor importance. We do find V sections with continuous outward grading that show only minor nicks, as one would expect for turbidity current valleys.

(7) The canyons tend to occur in localities favorable for the action of turbidity currents during the Ice Age. As off New England, many head at the break in slope, and some cut back into the shelf by headward erosion. Very few canyons start away down the slope or on deeply submerged ridges beyond the reach of glacial low sea levels.

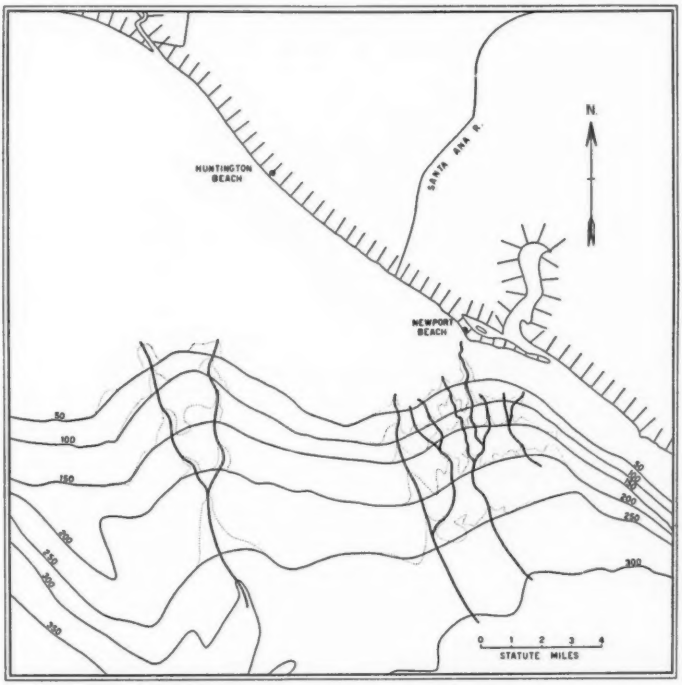

FIGURE 10. CHART OFF SOUTHERN CALIFORNIA
Showing canyons oblique to the coast but at right angles to the depth contours. Depths in fathoms.

In a number of canyons off river mouths a plentiful supply of fine sediment for setting up turbidity currents became available when the rivers deepened their channels during low sea levels.

Shepard (Shepard and Emery, 1941, p. 86) pointed out that many canyon heads are forked with the southern member approaching closer to the coast. "Furthermore in most of these cases the southern canyon is found directly north of either a point or of the termination of one of the short coast ranges." The author (1950, p. 516) invoked the action of oblique

northwesterly winds that brought to the heads of the canyons a plentiful supply of sediment for turbidity currents. Crowell (1952) concurs and gives much supporting evidence. But he emphasizes beach drifting rather than the conditions during the lowered glacial sea levels. This change in emphasis is unfortunate because, according to Shepard (1952, p. 92), the currents now pass around the heads. But that could not have occurred with an advanced shore line in glacial times. Nevertheless, in some cases Crowell appears to be right, and Shepard himself proved that sediments gather in canyon heads. Perhaps both points of view should be entertained, adding that some sediment will reach heads even where they are somewhat removed from the shore, especially during winter gales when the beaches are eroded vigorously.

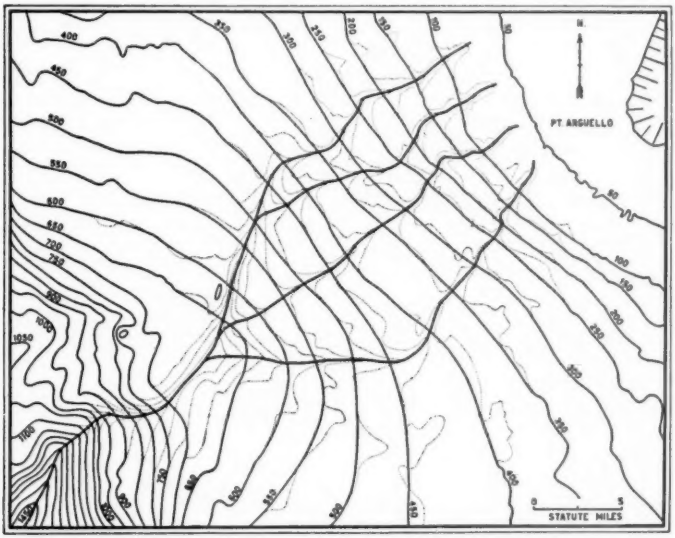

FIGURE 11. CHART SHOWING PARALLEL CANYONS ON FLAT SLOPE
Canyons converge to form Arguello Canyon in embayment of generalized contours. Depths in fathoms.

(8) Crowell (1952, p. 66) found that Redondo Canyon cuts into late Pleistocene deposits and that the head has been eroded since the rise of sea level following the low stand in the late Pleistocene. No major stream debouched into the canyon during or since the time of low sea level. A similar recent development is attributed to Newport Canyon.

According to Shepard, this is merely an example of the slumping effect he believes has been active since subsidence took place. It is remarkable that this postulated headward extension into recent sediment does not show up as a separate feature but fits smoothly onto the supposed ancient land valley farther out, (see also Shepard and Emery, 1941, Pl. 13). If submarine processes can develop a perfectly normal portion of a submarine canyon, why should they not be able to form similar parts farther out to sea? The objections to a dual nature of the walls, raised in connection with the New England canyons, also apply in these cases.

(9) Shepard (1948, p. 231; Emery *et al.*, 1952 p. 514) claims that several canyons show ancient deltas at their mouths. Slumping effects and nondepositional channels formed by turbidity currents have been added later. In the author's opinion there is no reason for expecting a drowned delta to be modified much by slumping. As much may be said for assuming that these fans were built by turbidity currents (and by slides?) from the coarse material dumped where the slope decreases and the confining influence of the canyon walls ceases,

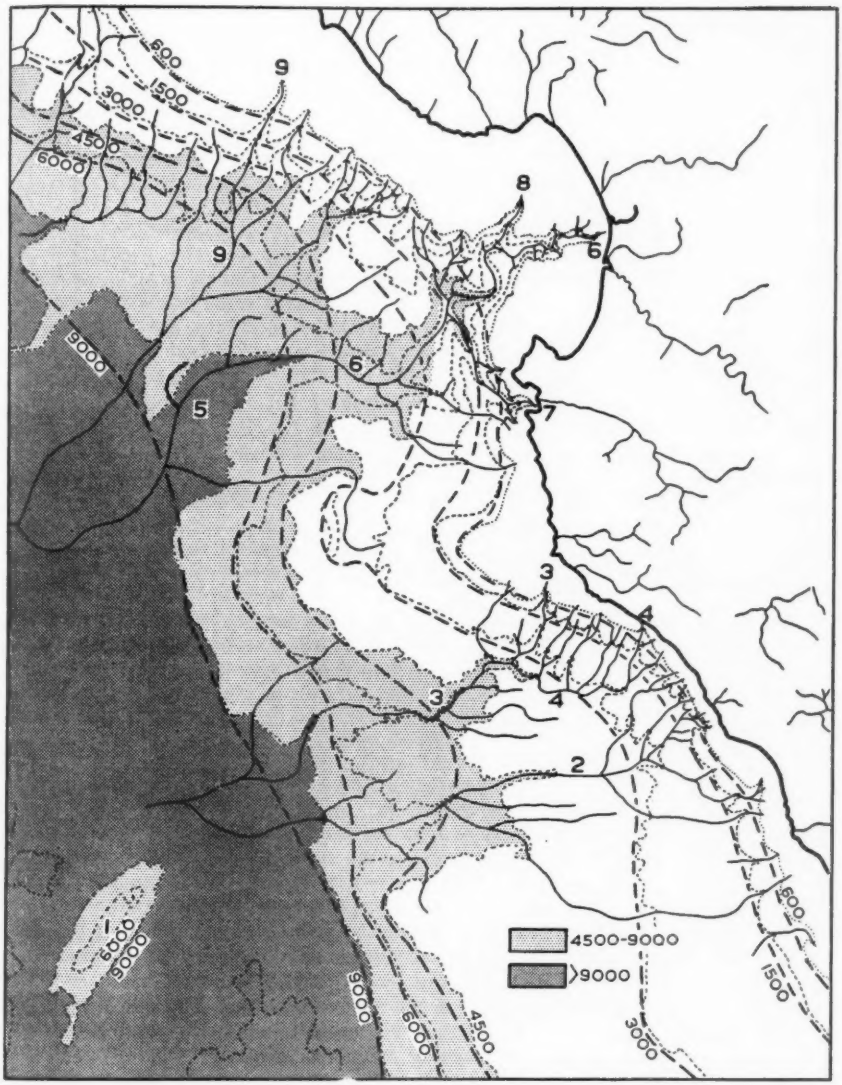

FIGURE 12. CHART OF DENDRITIC CANYONS OFF THE CALIFORNIA COAST

Inferred original shape of contours before canyons were cut is also shown. Data from chart by Shepard and Emery (1941, Chart II and III). Depths in feet. 1, Davidson Seamount; 2, Lucia Canyon; 3, Sur Canyon; 4, Partington Canyon; 5, Monterey Trough; 6, Monterey Canyon; 7, Carmel Canyon; 8, Soquel Canyon; 9, Ascension Canyon (After Kuenen, 1950, Fig. 215).

allowing the turbidity currents to fan out and slow down. The close similarity to the deltas in Swiss lakes that Shepard mentions points in the same direction, for, as far as shown by the published charts, the similarity is not with the horizontal subaerially built part of lake deltas, but with the subaqueous fore-set part on which underflow is active. True deltas may be incor-

porated in the low-angle cones fronting some of the California submarine canyons, but they appear to be smothered by later turbidity-current deposits like the New England deep-sea fans.

(11) The California canyons are more branching and winding, also steeper at the head than the New England type. A few start as narrow gorges cut entirely in hard rock. However, as a whole, they are remarkably similar

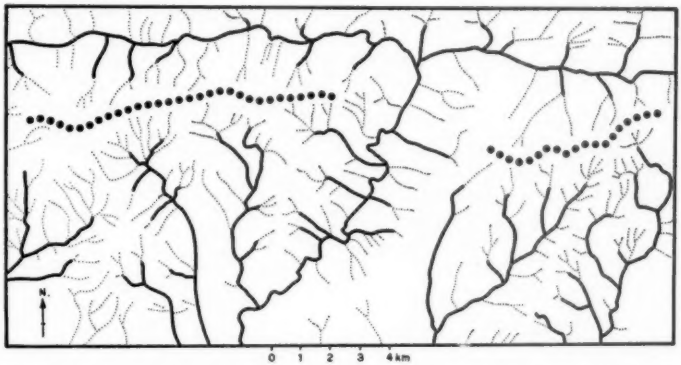

FIGURE 13. VALLEY SYSTEM AROUND SAN GABRIEL CANYON, NORTHWEST OF POMONA, CALIFORNIA
Even for major canyons and considering only the part south of the mountain range the system is more dendritic (less parallel) than with submarine canyons.

(10) Menard and Ludwick (1951) discovered levees along the lower reaches of some submarine canyons off California. Similar features have been charted by Shepard and occur along Swiss delta channels. They are also indicated along some of the canyons off the Rhone Delta (see also Buffington, 1952).

These levees must be depositional features and give important evidence of the size of the turbidity currents involved. The currents must have topped the levees to build them up. Sizes 20 meters deep by 800 meters wide are found in Lake Geneva; 100–150 by 3000 meters in the Mediterranean; 100–150 by 1000 meters off California. The size of these features demonstrates the vast scale on which turbidity flow acts. For example, with an effective density of 0.01, slope 1 in 40, and thickness of 20 meters, the velocity should be around 1½ meters per second (estimates for Lake Geneva). With density 0.2, slope 1 in 20, thickness 100 meters, the velocity would be 20 meters per second (estimates for marine canyons).

As far as known, levees are absent along the outer Hudson Canyon. As this valley is more than 500 meters deep turbidity currents could not be expected to spill over the edges.

to the New England canyons. The general shape of the longitudinal profiles is much the same. Bends are broadly rounded, and broad flat-bottomed stretches are absent. The slope of the continental terrace controls the direction. The depth of cutting is about equal. The most striking similarity is in the cross sections, which show the same narrow range of V-shaped profiles (Fig. 7). When these points of similarity are contrasted with the great variability of land valleys which range from narrow V-shaped canyons to broad, widely open troughs, it becomes obvious that the New England and California types must be attributed to the same causes. Hence, if the New England canyons were formed by turbidity currents, it is logical to assume the same process for the California canyons.

As turbidity currents could not cut hard rock, glacial scour must have been restricted to poorly consolidated materials. Therefore it only remains to show how soft-rock erosion could have produced canyons with occasional exposures of consolidated rock and a few narrow gorges in granite or cemented sediments.

Summary.—The following tentative picture is offered as a working hypothesis to those

better versed in the geology of California. Daly's theory gives a satisfactory explanation for the California canyons and differs slightly, but significantly from Shepard's latest proposal. Instead of assuming that drowned valleys were perpetuated by sliding and turbidity currents, which have no ability to erode, it is suggested that the ancient land surface was first smothered; later the poorly consolidated covering materials were eroded during the Ice Age, and to some extent in postglacial times to form the submarine canyons (Fig. 14).

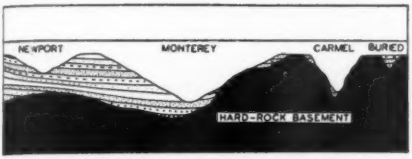

FIGURE 14. DIAGRAM ILLUSTRATING POSSIBLE CONSTITUTION OF DIFFERENT SUBMARINE CANYONS OFF CALIFORNIA

Some localities were particularly favorable to the generation of turbidity currents because of incompletely buried topographic depressions, local supply of sediment by rivers, and coastal configuration. Some narrow rocky land valleys were thus swept out (Carmel, Scripps, etc.), but the majority of old valleys may still lie buried in the terrace beneath sediments. Woodford is of opinion (personal communication) that there is little evidence of buried valleys in the terrace.

In some cases the turbidity currents only cleaned parts of the old valleys where these happened to offer small resistance. But other parts of these valleys did not conform to the requirement of following the present slope. Such parts remained buried.

Elsewhere a new valley cleaned off along its wall some small part of an ancient mountain slope, without conforming to the original drainage pattern. This may be the case for Monterey Canyon, which has granite overlain by sedimentary rock on one wall opposite a wall which has yielded only mud or soft sedimentary rock; or for Dume Canyon with basalt on the east side and mud with calcareous shale on the west.

A number of canyons are limited to poorly consolidated sediment deposited since the submergence. Some change in emphasis as compared to Daly's explanation is thus implied. The influence of the old, buried topography is thought to be limited, for only thus can one account for the marked similarity between the California and the New England canyons. The California group, as a whole, must reflect the action of turbidity currents guided only locally by the internal structure of the terrace. It is just this slight shifting of emphasis away from the old valley shapes and toward the natural outcome of turbidity-current activity which renders Daly's suggestion acceptable to anyone impressed by the insignificance of the differences in shape and size between the canyons off the west and east coasts of North America.

Shepard found that the canyons heading at the edge of the terrace have not been swept out recently, but that those approaching the shore receive much sediment which is periodically removed. The author attributed this to the favorable conditions in the latter type for the generation of turbidity currents (Kuenen, 1950, p. 514). Crowell emphasized the same aspect and gave evidence to show that this sweeping has even resulted in headward erosion in recent times.

Both the author and Shepard pointed out that any parts of the canyons above glacial sea levels may have been modified by streams during emergence (Kuenen, 1950, p. 525; Shepard, 1952, p. 86). Scripps Canyon, for instance, is the direct continuation of a small land canyon. Allowing for a slight recession of the cliffed coast since post-Glacial rise of sea level, the two must have joined in one uninterrupted system during low sea level.

The present working hypothesis has some advantage over Shepard's suggestion. The many points of similarity of the canyons in shape, size, pattern, and location with respect to the shelf edge on the east and west coasts of America can be attributed to identical modes of origin. The moderate differences of certain properties result from differences in local conditions. The preferred location of the canyons at localities favorable for the action of turbidity currents during the Ice Age (and, in part, at present), as against their absence off long stretches of the coast, coupled with the curious failure to cross the coast line, are easily

accounted for. The dependence on the dip of present submarine slopes, and recent extension of some heads are logical results of the proposed mechanism.

Exceptional Canyons

There are some exceptional canyons which require a different theory of origin. The Mississippi Trough-valley appears to have been largely influenced by sliding. As the Gulf Coast terrace is building up rapidly the trough cannot be old. Evidence is lacking in the Mississippi valley of a recent high stand of the land during which the river could have cut the trough subaerially. A submarine origin is therefore most probable.

The Congo Canyon enters the mouth of the estuary. Sediments from the river and the shelf are counteracted by some erosional process, probably turbidity currents. More field work is necessary to determine why this canyon has developed inside the coast line.

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REVISED CARBONATE-WATER ISOTOPIC TEMPERATURE SCALE

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ABSTRACT

The relationship between temperature and O^{18} content relative to that for a Cretaceous belemnite of the Pee Dee formation previously reported (Epstein, Buchsbaum, Lowenstam, and Urey, 1951) has been redetermined using modified procedures for removing organic matter from shells, and is found to be

$$t\,(°C) = 16.5 - 4.3\delta + 0.14\delta^2,$$

where δ is the difference in per mil of the O^{18} to O^{16} ratio between the sample and reference gas. The new relationship agrees with that determined by McCrea (1950) for inorganically precipitated calcium carbonate. Carbonate-carbon dioxide exchange experiments were done to determine the direct and indirect effects of organic matter in the shell on the mass spectrometer analyses.

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INTRODUCTION

Recently the writers (Epstein, Buchsbaum, Lowenstam, and Urey, 1951) reported on the determination of a temperature scale for measuring the temperature at which a marine shell-bearing animal grows its shell by determining the O^{18}/O^{16} ratio in the calcium carbonate of the shell. Although the method reported was fundamentally correct, extraneous oxygen was introduced into the calcium carbonate during one of the stages of the processing of the shell which introduced an error in the temperature scale. The complete problem was re-examined, and a modified method for preparing samples of carbon dioxide from the calcium carbonate of the shell is here described.

ACKNOWLEDGMENTS

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La Jolla, California, is gratefully acknowledged. A great deal of the laboratory work was done by Mrs. Toshiko Mayeda, and her efforts are appreciated.

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The work for this study was done at the Institute for Nuclear Studies at The University of Chicago.

Purification of the Calcium Carbonate of Marine Shells

General Discussion

The previously described purification of the calcareous shell consisted of heating 50 mgs of the powdered sample in a slow stream of helium at 470°C. The helium, which was purified from organic compounds by passing it over cupric oxide heated to 700°C and through a liquid nitrogen cooled trap, became contaminated with oxygen gas originating from the cupric oxide and from back diffusion through the furnace opening. The oxygen oxidized the organic material of the powdered shell to form CO_2 which in turn exchanged with the calcium carbonate of the shell. Since carbon dioxide formed with atmospheric oxygen analyzes approximately $-20‰$ relative to carbon dioxide extracted from calcium carbonate of a normal marine shell, the isotopic exchange lowered the O^{18}/O^{16} ratio of the oxygen of the calcium carbonate of the shells. The samples used for the temperature scale were treated in the same manner with the result that a near constant error was obtained. This approximate duplication masked the detection of the error, resulting in what was thought to be a valid temperature scale. Further work which showed a large discrepancy between the O^{18}/O^{16} analyses of the calcium carbonate in the prismatic and pearly layers taken from the same fragment of a shell led to the discovery of the error. An example of the discrepancies in the analyses is shown in Table 1.

$$\delta = (R_{sample} / R_{standard} - 1)\ 1000$$

where R_{sample} and $R_{standard}$ are ratios of $CO^{16}O^{18}$ to CO_2^{16} for the sample and standard reference gas respectively, and δ is essentially the difference between the O^{18} content of the sample and that of the reference gas.

Table 1.—O^{18} Analyses for CO_2 Samples from Different Layers of a Fragment from *Haliotis Rufescens* Grown at 21.0°C

Sample		δ	T_{calc}°C (old scale)
1.	Inner pearly layer	-3.6_9	$31._7$
2.	Intermediate layers	-3.0_7	$28._6$
3.		-2.0_2	$21._8$
4.		-1.7_6	$20._3$
5.	Outer prismatic layer	-1.8_9	$21._0$

All the analyses in Table 1 are erroneous. The differences in the above results we believe are due to a greater velocity of exchange of carbon dioxide with aragonite (pearly layer) as it is converted to calcite during heating. This is discussed further.

Purification Process

The process for the destruction of organic matter in the samples was modified. The apparatus is shown in Figure 1. Helium gas flowing at a rate of 0.4 cc per second passes through a copper-filled furnace kept at approximately 500°C, through an activated charcoal-filled trap cooled with liquid nitrogen, and into the roasting furnace. The continuous flow of helium sweeps the volatile decomposition products of the heated organic compounds away from the calcium carbonate and at the same time provides an inert atmosphere over the sample. The sample, in a platinum boat, is inserted into the furnace while the temperature of the furnace is low (less than 200°C) to permit sweeping out of all the air from the furnace before heating of the sample begins. The opening of the furnace is covered by a cap containing a capillary opening to prevent back diffusion of air. Sweeping with helium continues for 20 minutes before the heater of the furnace is turned on. This is sufficient to sweep out the furnace about six times. The sample roasts in the furnace at least 30 minutes after the temperature of the

furnace reaches 470°C. The resulting calcium carbonate, which is gray, is then reacted with phosphoric acid to release the carbon dioxide

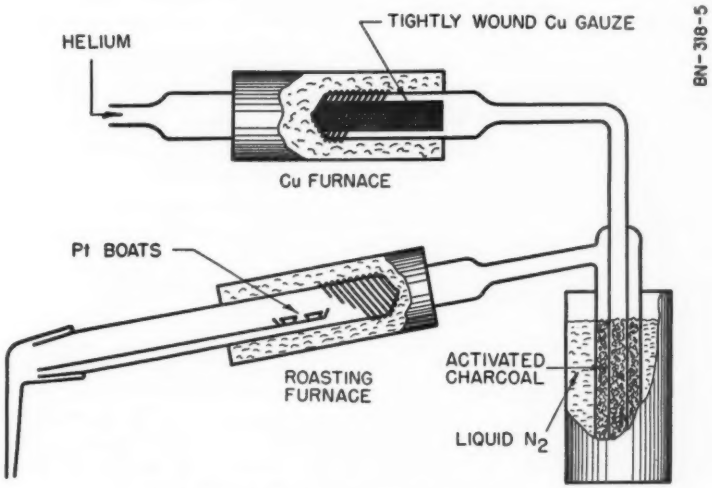

FIGURE 1.—PURIFICATION APPARATUS

for mass spectrometric analysis in the manner described in the previous publication. Under these conditions of purification the size of the sample, the length of time beyond 30 minutes of heating, and the difference in the type of layer in the abalone shell do not affect the results of the analyses, as shown in Table 2. The two fragments grew at different temperatures, and hence the analyses should be different.

The effect of heating by the modified procedure is to decrease the O^{18}/O^{16} ratio of the oxygen in the resulting calcium carbonate, but to a lesser degree than in the case of heating in an atmosphere of a mixture of helium and oxygen. The fact that roasting is still necessary is shown in Table 3.

The data show that heating affects the isotopic analysis of samples from different animals differently. Although the first three samples listed were grown simultaneously in the same tank, their respective mass spectrometer analyses check only after the heating procedure has been followed. The analysis for the *Strombus* checks those of the other specimen only because the difference in O^{18}/O^{16} ratios between the waters in which the *Strombus* and the other specimens grew is 1.6‰, which is nearly sufficient to compensate for the difference in the temperature at which the specimen grew.

TABLE 2.—EFFECT OF VARYING CONDITIONS OF HEATING ON THE O^{18} CONTENT OF $CaCO_3$ FROM *Haliotis* SAMPLES

Sample	Weight mgs	Time of heating at 470°C min.	δ
Fragment I			
Pearly layer	10	30	-1.1_9, -1.4_4
	50	30	-1.3_3, -1.2_8, -1.0_0
Prismatic layer	50	30	-1.2_7
Fragment II			
Pearly layer	50	30	-0.8_8
	50	75	-0.8_6

Isotopic Exchange Experiments

Although the heating of the samples in helium gas with rigorous exclusion of oxygen and other precautions outlined have given very consistent results, the investigation of the cause for the

previous errors is important if we are to be confident that the present consistency in results is not due to systematic errors. In the former paper, it was concluded that the reaction of shells containing organic matter with phosphoric acid does not produce an impurity of masses 44 and 46 which interferes with the mass spectrometric analyses, and we know of no reason to change this view.

TABLE 3.—EFFECTS OF ROASTING ON ANALYSIS OF O^{18}/O^{16} RATIO OF $CaCO_3$ FROM MARINE SHELLS

Sample	Growth Temp. °C	Fresh unheated sample δ	Sample heated in He gas δ	Change due to heating
Snail (*Kelletia*)...	21.5	-0.3_3	-1.4_3	-1.1_0
Haliotis (prismatic layer).....	21.5	-0.9_6	-1.5_1	-0.5_5
Calcareous worms	21.5	-0.4_6	-1.3_6	-0.9_0
Snail (*Strombus*)..	29.5	-1.3	-1.4_5	-0.1_5

However, organic matter may affect the O^{18} content of the carbon dioxide extracted from shell in the following ways: (1) Its reaction with phosphoric acid may produce carbon dioxide different in isotopic composition from that produced from calcium carbonate of the shell; (2) Even in the absence of oxygen during the heating of shell sample in helium at 470°C, carbon dioxide or other oxygen compounds produced from the decomposition of shell organic matter may exchange with the carbonate.

These possibilities could be investigated if the oxygen in the organic matter could be labeled and its contribution to the carbon dioxide extracted from the powdered shell detected. It was hoped that the oxygen in the organic matter could be preferentially enriched in O^{18} relative to the $CaCO_3$ of the shell (and thus labeled) by heating the powder in an atmosphere of carbon dioxide enriched in O^{18} at temperatures high enough to facilitate exchange but low enough so that all the organic matter is not driven from the powder. Such preferential O^{18} enrichment of the organic material should be detected by comparing the mass spectrometer analyses for carbon dioxide samples extracted from the above powder with that extracted from powder which has been subjected to the above exchange after its organic matter is removed by heating at 470°C in an atmosphere of helium. The procedure used for these exchange experiments is described below.

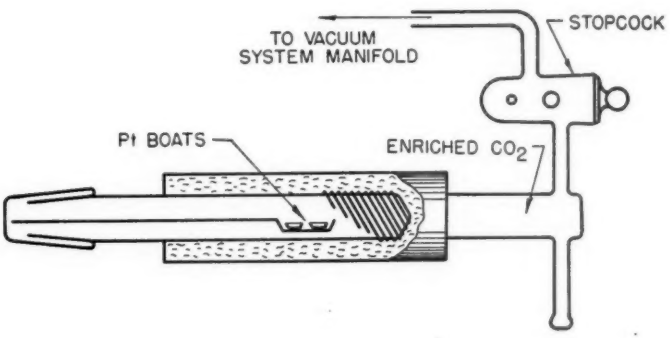

FIGURE 2.—FURNACE FOR EXCHANGE EXPERIMENTS

A fragment of shell from a *Strombus gigas* which was 100 per cent aragonite and a fragment from an oyster shell were separately powdered, and a 30-milligram aliquot from each powder sample heated in a helium gas stream by the method described in the purification procedure. This procedure converted the aragonite to calcite in the former case and

destroyed the organic matter in both cases. Thirty milligrams, which will be designated as Sample I of the original unheated aragonite powdered shell, and an equal amount of the heated powder, designated as Sample II, were placed simultaneously in separate boats in a furnace shown in Figure 2. The system was thoroughly evacuated, after which 85 cc of carbon dioxide which analyzed +100‰ higher in O^{18} relative to our standard was introduced into the furnace. The resulting pressure in the furnace was approximately half an atmosphere. The temperature of the furnace was raised to the desired value, and the samples allowed to exchange with the carbon dioxide for the desired length of time. The O^{18} enriched carbon dioxide was then removed, and 15 milligrams of each sample was reacted with phosphoric acid as previously described. The same procedure was repeated using the powdered oyster shell with the untreated sample designated as Sample III and the helium heated sample as Sample IV. The mass spectrometric analyses of the gas samples are given in Table 4. The crystalline form of the calcium carbonates was determined by x-ray analyses.

Comparison of the difference in analyses between Sample I(a) and Sample I(c) with the difference in analysis between Sample II(a) and Sample II(c) shows that in the case of Sample I, which contained organic matter, exchange with O^{18} enriched carbon dioxide resulted in an enrichment of the O^{18} of the carbon dioxide extracted from the shell of 1.7_9 while in the case of Sample II, from which organic matter was first removed before exchange with O^{18} enriched carbon dioxide was permitted to proceed, the resultant enrichment in O^{18} was 0.4_2. Thus organic matter must directly or indirectly affect the isotopic composition of the carbon dioxide extracted from the shell. Since for Sample III the enrichment due to exchange at 320°C [comparison between III(a) and III(b)] is 1.64 and is similar to that for Sample I it appears that in this case aragonite and calcite powders behave similarly to exchange with carbon dioxide. The analyses for Sample I(d) show a very marked increase in O^{18} when aragonite transformed to calcite, whereas such an increase did not occur for the Sample II(d) which was treated similarly. Although further work is necessary to understand this effect, it is very likely that such transformation should weaken the bonds in the calcium carbonate and lower the activation energy necessary for exchange between calcium carbonate and carbon dioxide. This increase in exchange rate agrees with the results of Table 1 —namely, the aragonite pearly layer of an abalone shell was more markedly affected by surrounding carbon dioxide than was the calcite prismatic layer when both were heated at 470°C.

TABLE 4.—POWDERED SHELL-CO_2 EXCHANGE

Temp. of heating °C	Time of heating Mins.	Sample I (Crystal form after exchange) $\delta^0/_{00}$	Sample II Calcite $\delta^0/_{00}$
Not heated		(a) $+0.1_9$ (100% aragonite)	(a) -0.2_1
320	30	(b) $+1.3_0$ (100% aragonite)	(b) $+0.0_5$
320	60	(c) $+1.9_8$ (100% aragonite)	(c) $+0.2_4$
420	60	(d) $+14.5_5$ (80% calcite)	(d) $+0.7_7$
		Sample III Calcite	Sample IV Calcite
Not heated		(a) -2.0_2	(a) -2.3_3
320	60	(b) -0.3_8	(b) -1.5_8
375	60	(c) $+1.9_4$	

TABLE 5.—EFFECT OF HEATING AT 475°C IN HELIUM ON O^{18} EXCHANGED SAMPLES

Sample (Exchange temp)	Before heating in Helium (from Table 4) δ	After heating in Helium δ	Change in O^{18} content δ
I(c) (320°C)	1.98	1.20	−0.78
III(b) (320°C)	−0.38	-1.0_9	−0.71
III(c) (375°C)	1.94	1.1_8	−0.76
IV(b) (375°C)	−1.58	-1.9_1	−0.33

If the powdered samples of shell which were not first heated at 470°C and which were subjected to exchange with O^{18} enriched carbon dioxide contain preferentially O^{18} enriched organic compounds, then further heating of these samples in an atmosphere of helium at 470°C should destroy the organic matter, resulting in a lowering of the O^{18} content of the carbon dioxide extracted from such heated samples. Accordingly, aliquots from Samples I(c), III(b), III(c), and IV(b) were heated at 470°C, and the carbon dioxide from these samples analyzed. The results are shown in Table 5.

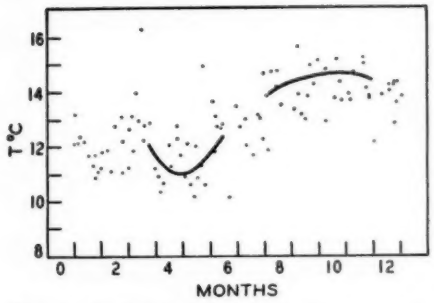

FIGURE 3.—SURFACE OCEAN TEMPERATURES AT PACIFIC GROVE, CALIFORNIA
Average maximum 14.5°C, Average minimum 11.0°C.

Table 5 shows that the first three samples listed have lost O^{18} to a greater extent than did the last sample during the heating at 475°C. The first three samples contained organic matter while exchanging with O^{18} enriched carbon dioxide, while the last sample was purified of organic matter before it was allowed to exchange. It seems, then, that heating at 470°C in an atmosphere of helium eliminates some organic matter which would otherwise affect the analyses of the carbon dioxide extracted from a powdered shell. That no exchange has taken place during this 470°C heating is shown by the fact that Samples I(c) and III(b) have been reduced in O^{18} by the same amount. Since the former sample was aragonite which converted to calcite at 470°C and the latter sample was originally calcite and thus suffered no crystallographic transformation during the

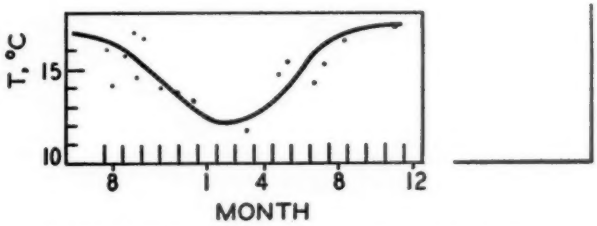

FIGURE 4.—SURFACE OCEAN TEMPERATURES AT SANTO TOMAS
Average maximum 17°C, Average minimum 12°C.

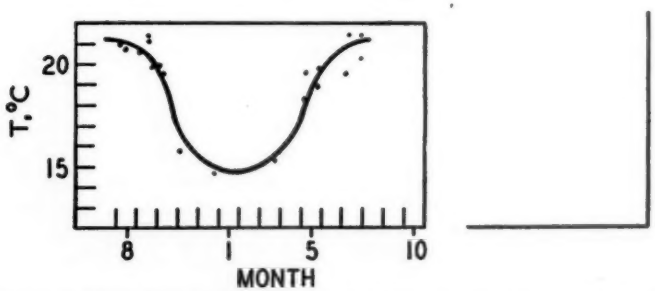

FIGURE 5.—SURFACE OCEAN TEMPERATURES AT POINT BETWEEN MORRO AND ENSENADA
Average maximum 21°C, Average minimum 15°C.

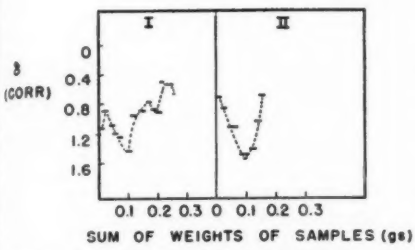

FIGURE 6.—PLOT OF O^{18} CONCENTRATION VS. GROWTH SEQUENCE OF LAYERS IN ABALONE SHELL

(*Haliotis*, Pacific Grove, Temperature 11–14.5°C)

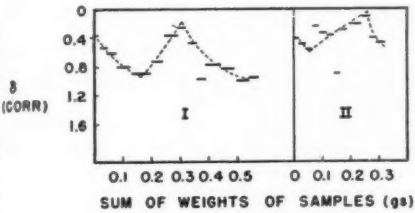

FIGURE 7.—PLOT OF O^{18} CONCENTRATION VS. GROWTH SEQUENCE OF LAYERS IN ABALONE SHELL

(*Haliotis*, St. Tomas, Temperature 12–17°C)

enrichments are particularly large in the case of the sample that had organic matter during the exchange experiments. That calcium carbonate should exchange more readily when intimately mixed with decomposing organic matter is not surprising in the light of the work of Armstrong and Schubert (1947). They found that presence of water increases the rate of exchange between CO_2 and carbonate. Possibly the less marked exchange that occurred in the case of the shell powder which had been previously purified of organic matter is due to water from its neighboring unpurified sample which had contaminated carbon dioxide in the furnace, and that the apparent loss of O^{18} by Sample IV(b) after being heated in helium at 470°C is due to loss of organic matter picked up by this sample from its neighboring Sample III(b) during the exchange experiment.

It has been noted that finely powdered Foraminifera which have been exposed to air for a few months will in some cases show a lower O^{18} content relative to that of freshly powdered shell. Fortunately this effect does not seem to exist for the samples used in the determination of the temperature scale since some of the

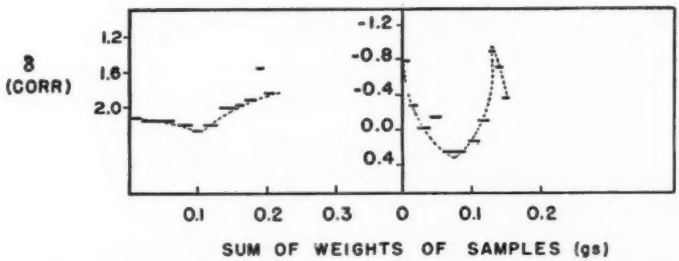

FIGURE 8.—PLOT OF O^{18} CONCENTRATION VS. GROWTH SEQUENCE OF LAYERS IN *Macoma* sp. (at left; Middle Channel, Temperature 7–9°C) AND IN ABALONE (at right, *Haliotis*; Point between Morro and Ensenada points, Temperature 15–21°C)

heating, then if exchange had taken place during the heating Sample I(c) would have been less depleted in O^{18} than Sample III(b).

The results in Column 3, when compared with the O^{18} analyses for the samples of oyster and *Strombus* which have merely been heated in helium at 470°C (IV(a) and II(a) respectively) show that there is a resultant enrichment in O^{18} in the calcium carbonate that occurred during the exchange experiment. The results reported here have been obtained over a period of a year. This effect should be guarded against in any further work on freshly gathered sea shells.

TEMPERATURE SCALE

With the development of a reliable process for destroying organic matter in organically precipitated calcium carbonate the work on the temperature scale was continued.

TABLE 6.—DESCRIPTION OF SHELL SAMPLE USED FOR DETERMINATION OF TEMPERATURE SCALE

Location	Specimen	Part of specimen used	Temp. °C	Water Analysis O^{18}
Thermostated tank at Hopkins Marine Station, Pacific Grove, Calif.	1. Black abalone, *Haliotis cracherodii*	Regenerated edge (mother of pearl)	19	−0.44
	2. Black abalone, *Haliotis cracherodii*	Regenerated holes of spec. 1 + spec. 2 (mother of pearl)		
	3. Black abalone, *Haliotis cracherodii*	Regenerated hole (mother of pearl)		
	4. Black abalone, *Haliotis cracherodii*	Regenerated notch in edge of shell		
	5. Black abalone, *Haliotis cracherodii*	Regenerated notch		
	6. Black abalone, *Haliotis cracherodii*	Regenerated notch		
	7. Black abalone, *Haliotis cracherodii*	Regenerated notch		
Thermostated tank at Scripps Institution of Oceanography, La Jolla, Calif.	8. Black abalone, *Haliotis cracherodii*	Regenerated hole	21.5	−0.45
	9 (a). Black abalone, *Haliotis cracherodii*	Pearly layer of regenerated notch		
	(b). Black abalone, *Haliotis cracherodii*	Prismatic layer of regenerated notch		
	10 (a). Black abalone, *Haliotis cracherodii*	Pearly layer of regenerated notch		
	(b). Black abalone, *Haliotis cracherodii*	Prismatic layer of the same		
	11 (a). Black abalone, *Haliotis cracherodii*	Pearly layer of regenerated notch		
	(b). Black abalone, *Haliotis cracherodii*	Prismatic layer of the same		
	12. Black abalone, *Haliotis cracherodii*	Average of regenerated notch		
	13 (a). Black abalone, *Haliotis cracherodii*	Pearly layer of regenerated notch		
	(b). Black abalone, *Haliotis cracherodii*	Prismatic layer of the same		
	14. Snail *Kellatia kellettii*	Average sample of regenerated edge		
	15. Oyster Mytelles *californianus*	Average sample of regenerated edge		
	16. Calcareous worm	Average sample		
Bermuda, tank in laboratory	17. *Strombus gigas*	Regenerated edge	29.5	+1.13
Hopkins Marine Station	18. *Haliotis cracherodii*	Fragment used in seasonal variation	Min. 11 Max. 14.5	−0.44

TABLE 6—*Continued*

Location	Specimen	Part of specimen used	Temp. °C.	Water Analysis O^{18}
Santo Tomas	19. *Haliotis cracherodii*	Fragment used in seasonal variation	Min. 12 Max. 17	−0.43
Point between Morro and Ensenada Points	20. *Haliotis rufescens*	Fragment used in seasonal variation	Min. 15 Max. 21	−0.29
Middle Channel, Puget Sound	21. *Macoma* sp.	Fragment used in seasonal variation	Min. 7.2 Max. 9.5	−0.63

To establish the relationship between O^{18}/O^{16} ratio of the oxygen in calcium carbonate precipitated by marine shells and the temperature at which they were precipitated, regenerated fragments of shell grown in temperature-controlled tanks as well as shells collected in their natural surroundings were used. One group of animals whose shells were notched or drilled and which then repaired their shells in a thermostated tank at 19°C was sent to us by Dr. Rolf Bolin from the Hopkins Marine Laboratory. A *Strombus gigas* which regenerated part of its shell at 29.5°C in a tank at Bermuda provided shell for a high-temperature calibration point. A third group of animals which regenerated their notched and drilled shells at 21.5°C was sent to us by Dr. Carl Hubbs. These three groups then provided three calibration points.

The other points on the curve were obtained by determining the variation of O^{18}/O^{16} ratio in calcium carbonate samples ground off from successive growth layers from a rectangular piece of shell. The marine shells grew in a location whose seasonal variation of temperature is known. The locations chosen were off the shores of Santo Tomas, a point between Morro and Ensenada points on the west coast of Lower California, and at Pacific Grove, California, and from Middle Channel in Friday Harbour, Puget Sound, Washington. The seasonal temperature variation of the first three locations is shown in Figures 3, 4, and 5. The data for the first two localities mentioned were sent to us by Dr. Carl L. Hubbs, and the data for the curve in Figure 5 are from Skagsberg and Phelps (1946). The samples from Middle

TABLE 7.—ISOTOPIC ANALYSES OF ORGANIC CALCIUM CARBONATE SAMPLES

Spec. No. (Table 6)	δ (obs.)	δ Average	δ Average corrected	Temp. °C
1	−1.03	−1.03	−0.59	19
2	−1.28	−1.28	−0.84	19
3	−1.20	−1.20	−0.76	19
4	−1.06	−1.06	−0.62	19
5	−1.06	−1.06	−0.62	19
6	−1.38, −1.10	−1.24	−0.80	19
7	−1.12, −0.88	−1.00	−0.56	19
8	−2.23, −1.80, −2.23	−2.09	−1.65	21.5
9(a)	−1.50, −1.50	−1.50	−1.05	21.5
(b)	−1.63, −1.30, −1.64, −1.52, −1.39			
10(a)	−1.57, −1.64	−1.47	−1.02	21.5
(b)	−1.22, −1.43			
11(a)	−1.46, −1.64, −1.64	−1.44	−0.99	21.5
(b)	−1.12			
12	−1.52	−1.52	−1.07	21.5
13(a)	−1.70, −1.78, −1.86, −1.60	−1.64	−1.19	21.5
(b)	−1.42, −1.50			
14(a)	−1.48, −1.43, −1.27	−1.46	−1.01	21.5
(b)	−1.53, −1.45, −1.62			
15	−1.44, −1.52	−1.48	−1.03	21.5
16	−1.29, −1.42	−1.36	−0.91	21.5
17(a)	−1.41, −1.40, −1.50	−1.46	−2.59	29.5
(b)	−1.51			

TABLE 7—*Continued*

Spec. No. (Table 6)	δ (obs.)	δ Average	δ Average corrected	Temp. °C
18(a)		1.02	1.46	11
(b)		0.06	0.50	14.5
19(a)		0.58	1.01	12
(b)		−0.35	0.08	17
20(a)		0.03	0.32	15
(b)		−1.20	−0.91	21
21(a)		1.64	2.27	7
(b)		1.20	1.83	9

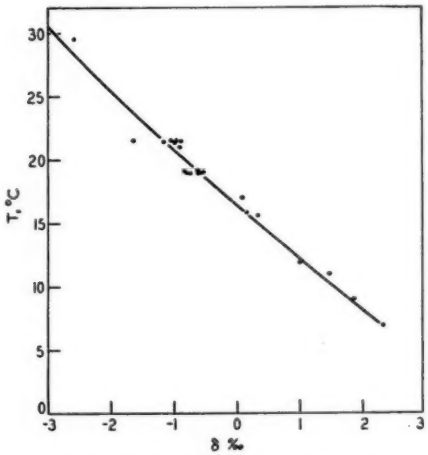

FIGURE 9.—ISOTOPIC TEMPERATURE SCALE

Channel locality lived at a depth of 145 meters, and the temperature ranges from about 9°C during July to about 7°C during December. The average maxima and minima of the temperatures of the localities were then estimated and matched with the maxima and minima in the respective curves resulting from the plot against the succession of layers in the shells shown in Figures 6, 7, and 8.

Table 6 lists the species and their location and temperatures of shell growth. The isotopic composition of the water in which these forms grew has been more carefully redetermined in conjunction with a complete program of determining the variation of isotopic composition waters. From this study which will be reported elsewhere the analysis for the O^{18}/O^{16} ratio of carbon dioxide equilibrated with the average marine water was found to be 0.0‰ relative to our working standard gas. Therefore the correction, due to differences of O^{18}/O^{16} of water in which the marine animals grow, of the isotopic analysis of $CaCO_3$ from the average isotopic composition of the oceans will be the negative value of the analyses for the water made in the same manner and is listed in the last column of Table 6. In Table 7 are listed the specimen numbers as specified in Table 6, the δ analyses, and the δ analyses corrected for the isotopic composition of the water. Figure 9 shows a plot of the results of the last two columns of Table 7.

Least squares calculations of the data in Table 7 gave the following equation:

$$t = 16.5 - 4.3\delta + 0.14\delta^2,$$

where t is the temperature in °C, δ is the permil differences between the ratio of masses 46 and 44 of the sample and of the working gas. The intercept 16.5 is characteristic of the standard and the choice of the isotopic composition of the average marine waters. The standard deviation for t is equal to ±0.6°C.

DISCUSSION AND CONCLUSIONS

There is one experimental point on the curve which deviates far beyond the experimental error from the calculated curve. This point was determined using mother of pearl shell which was regenerated at 21.5°C by the abalone in the process of covering a hole drilled in the shell. Similar points at 19°C deviate in the same direction, to a much lesser degree. Possibly such calcium carbonate may have been laid down by the abalone with sufficient rapidity to prevent complete isotopic equilibration between the solid calcium carbonate and surrounding water.

Least-squares calculations of data determined by McCrea (1950) for inorganically precipitated calcium carbonate in the range of 7–25°C give slopes of 5.2 and 4.1 and intercepts of 17.5 and 13.0 for series using Florida and

Cape Cod waters respectively. The value of the slopes is essentially in agreement with the present value of 4.3. Florida water of salinity 36.5 collected at Jacksonville analyzes +0.7‰, while the standard used by McCrea analyzes +0.4‰ relative to our present working standard. The intercept obtained by McCrea where he used Florida water, when corrected for the differences in the δ of the water and standards, becomes 16.0. The close agreement between the temperature scales based on inorganic and organic precipitation leads us to conclude that a valid isotopic temperature scale is here reported. The second equation from McCrea's data cannot be corrected since we do not have a sample of the water which he used.

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